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THE LAITAURE DELTA

A STUDY OF DELTAIC MORPHOLOGY AND PROCESSES

BY

VALTER AXELSSON

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Inaugural Dissertation

BY

VALTER AXELSSON

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BY
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With two plates

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PREFACE

The extensive delta formations which often arise at river mouths in lakes and seas are an obvious testimony to the eroding, transporting and accumulating action of the rivers and rapidly reflect changes in the activity of the fluviatile, marine and/or lacustrine processes. Continual splitting, formation, oscillation and closing of distributaries characterize the active phase of delta formation, by which the normally branching delta pattern is created and modified.

This study deals especially with modern, gently sloping, subaquatic-subaerial types of delta, and the morphological delta pattern is considered principally from fluviatile points of view. The aim of this work has been on the one hand to give a survey of recent advances in the study of deltaic morphology and processes, and on the other to examplify and further elucidate certain deltaic features by a description of a rapidly growing delta.

The first part of this work deals with certain general conditions for the formation of deltas, the flow pattern at the mouths of deltaic channels, some characteristic morphological features of the delta front and the delta plain, the splitting and closing of distributaries, and with the sedimentary structures of various deltaic deposits.

The morphological development of a branching river pattern is greatly dependent on the special conditions for erosion, transportation and deposition of sediment at the head of channel forks. These processes, that are of fundamental importance for the development of deltas, are further discussed in the second part of this work. This part also presents the results of an experimental study on the diversion of bed load at a channel fork, performed at the Geomorphological Laboratory in Uppsala.

The third, and final, part is devoted to an account of hydrological, morphological and sedimentological *data* of the Laitaure delta. The field works carried out by the author on this delta were started in 1953 and ended in 1959, and are partly presented in some preliminary reports.

The Laitaure delta has an area of only 10 km^2 but is nevertheless one of the largest and most rapidly growing lake deltas in Sweden. It is situated at the eastern border of the Scandinavian mountain range, just north of the Arctic Circle, and has been formed in Lake Laitaure by the river Rapaälven, which drains the central part of the high mountain region of Sarek.

The drainage basin upstream of the Laitaure delta has an area of 684 km^2 and is part of the Sarek national park, which comprises the most glacierized high mountain region of Sweden. About 40 glaciers, having at present a total area of about 80 km^2, are drained by tributaries to Rapaälven. Glaciers thus constitute about 12 per cent of the drainage area upstream of the Laitaure delta. The larger part of the drainage basin is highly inaccessible, and is still rather poorly known both geographically and geologically. Most of our knowledge about this drainage basin comes from the scientific investigations, conducted by *Axel Hamberg*, who spent practically every summer during the years 1895—1931 in Sarek.

Due to the absence of larger lakes, the steep topography and the small thickness of the soil cover within the drainage basin, the water level in Rapaälven and in the deltaic distributaries rises and falls very rapidly in summer with changes in precipitation and temperature. During dry periods in summer with a marked daily variation in air temperature, there is also a very distinct daily variation in water stage, as well as in water discharge. The annual variation in water discharge is very great. At the proximal part of the delta the mean annual extreme values amount to 294 m^3/sec. and 0.6 m^3/sec. respectively.

At the proximal part of the delta the discharge of sediment amounts, on an average, to about 185,000 metric tons/year. This value of the sediment discharge is very high in comparison with most rivers in Sweden. About 14 per cent of this amount is transported as bed load. The annual variation in sediment discharge is extremely great. The mean annual extreme values are lower than 0.01 ton/day and higher than 20,000 tons/day respectively. Half of the total annual discharge of sediment takes place, on an average, during a total period of about 5 days. This means that some years half of the total annual delta growth may take place during a single week.

Lake Laitaure formerly had a length of 15 km but is nowadays half filled up by the Laitaure delta. Physiographically this delta has in the

main the normal appearance with numerous active and abandoned river channels, more or less marked levee ridges, and shallow inter-levee basins of different types. The shape of the delta plain, however, is not lobate but elongate, because its lateral extent is limited by the small width of the original lake basin. Initial topographic irregularities influence the splitting and closing of channels and the formation of central and marginal delta basins, but in the main the deltaic pattern is the result of fluvial activity.

The average longitudinal slope of the delta plain, as defined by the average slope of the levee crests along the active channels, is about 0.2 per mille in the middle part of the delta, and about 0.3 per mille in the proximal part. The channels are shallow. In winter most of the smaller channels freeze to the bottom. The levees have comparatively steep backslopes. The larger inter-levee basins are subdivided by minor levee complexes into smaller lake basins. Several of the delta lakes are partly filled up by miniature deltas, formed at the mouths of inflowing channels.

At the delta front the levees are low and poorly developed, the channels are poorly defined and bar deposits of active and abandoned channels merge laterally to form greater distal lobes. The vertical and lateral extent of steeply inclined delta-front slopes of a real foreset character is small at the present delta front in Laitaure.

The rate of advance of the Laitaure delta is high as compared to that of most lake deltas in Sweden. If we assume that the amount of sediment transported to the Laitaure delta will be of the same order in the future as at present the remaining part of Laitaure will be filled up by sediments within a period of 1,000 years.

For the sake of simplicity the symbols used in this paper are listed and defined at the end of Part III. They are also defined in each of the three parts where they first appear.

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VALTER AXELSSON

Part I

DELTAIC MORPHOLOGY AND PROCESSES

Chapter 1

GENERAL

The term "delta"

The term "delta" goes back to *Herodotus* (c. 484—425 B. C.), who compared the more or less triangular shape of the alluvial land enclosed between the diverging distributaries of the Nile and the sea to that of the Greek letter Δ. Later on, the word delta was applied as the general term for "alluvial land, formed by a river at its mouth, without reference to its precise shape" (*Lyell*, 1854).

The most frequently cited definition of the term "delta" is probably that of *Barrell* (1912, p. 381). He defined a delta as "a deposit partly subaerial built by a river into or against a body of permanent water". In order that this definition may also cover ancient deltas, *Barrell's* "body of permanent water" should, as *Nevin and Trainer* (1927, p. 454) suggest, be replaced by the expression "bodies of standing water", because many bodies of water must be regarded as ephemeral. *Barrell* also makes what is from the definition point of view a very important observation, namely, that "a delta consists of a combination of terrestrial and marine, or at least lacustrine strata, and differs from other modes of sedimentation in this respect" (p. 381).

Bates (1953), who exploited the jet analogy in his discussion of deltaic formation, defines a delta as "a deposit built by jet flow into or within a permanent body of water" (p. 2125). *Bates* thus uses the term "delta" in a more extensive sense and includes in it permanently submerged deltaic deposits. However, his definition cannot be considered satisfactory, as not all subaqueous deposits formed by jet flow, such as, for instance, flat bottomset beds, have a real delta character. Besides, as *van Straaten* (1960, p. 412) points out, "an essential part of river delta sediments, both subaqueous and subaerial, are formed by other processes than jet flow".

In general the term "delta" is used in a rather limited sense. *Throwbridge* (1930, p. 868) considers that "the noun delta and the adjective deltaic should be applied only to those sediments that are deposited by rivers under conditions that exist at and near their mouths". In Fennoscandia, where especially the ancient glacifluvial formations have attracted great interest, the delta concept has often been used in a very extensive sense and also come to include delta-like accumulation forms built up on the land surface. For the latter the term "supraaquatic (supraaqueous, supramarine, subaerial) deltas" has been used, while the deltas formed in lakes and seas have been designated "subaquatic (subaqueous, submarine)".

The supraaquatic deltas are usually built up by relatively rapidly oscillating channels. According to *Hjulström* (1952, 1953, 1958) the supraaquatic deltas differ from the subaquatic deltas owing to their greater slope and different stratigraphic structure.

These supraaquatic formations, in common with some river-mouth deltas, have no real foreset beds, but they differ from all river-mouth deltas by having no bottomset beds. A steep slope and a relatively rapidly oscillating braided-river system also characterize especially the more proximal parts of many glacifluvial river-mouth deltas and likewise the coarse-grained accumulations often formed at the mouths of mountain torrents. The supraaquatic part of these deltas often has the appearance of an alluvial fan. From the morphological point of view, these deltas may therefore be regarded as a transitional form between steep sloping, supraaquatic deltas and gently sloping, river-mouth deltas. To those alluvial deposits that are built up where swift, muddy tributaries enter slow-flowing main streams, the term "tributary river deltas" has

sometimes been applied. In the international literature, however, modern and fossil real supraaquatic braided-river accumulations are generally not called deltas but are given special names, such as alluvial cones or fans, outwash plains or sandurs, etc.

On many river-mouth deltas the levee crests are built up to normal high-water level and therefore relatively large areas will be exposed at low water. These subaquatic-supraaquatic deltas could therefore also be characterized as subaquatic-subaerial, as distinguished from other real subaquatic (or submarine) and permanently submerged deltas, such as occur, for instance, in estuarine environments.

Subaquatic, permanently submerged deltas also occur at great depths, namely, at the mouths of submarine canyons. A collation of observations of such formations has been made by *Bates, Mooney and Bershad* (1959). For these deltas they have suggested the term "submarine canyon deltas" (pp. 596—597).

A separate subaquatic type of delta, almost purely marine in origin, is represented by tidal deltas. The term is used for certain accumulations of sediment produced by tidal currents on one or both sides of the inlets of tidal lagoons. See, for instance, *Davis* (1899), *D. W. Johnson* (1919), *Kuenen* (1950) and *Guilcher* (1954). Detailed descriptions of such formations have been given by *van Veen* (1936, 1950).

Whether the term "delta" is used in a narrow or in a wide sense, the word delta without attribute should only be used for subaquatic-subaerial types of delta. When the word delta is used for certain other accumulation forms of a delta character or in a more extensive sense, it should be preceded by various attributes, such as supraaquatic, subaquatic, submarine, submerged, tidal, etc.

The present paper deals only with modern, gently sloping, subaquatic-subaerial types of delta.

Deltaic formation

In courses of running water there is a continuous transport and re-deposition of particles detached by weathering and erosion. The finer-grained particles (fine sand, silt and clay) are transported in suspension, while the coarse sand and heavier particles are mainly carried along the bottom. If the current velocity decreases, the transport capacity will also decrease and the transported material will be deposited. At river mouths in lakes and seas there is usually a rapid decrease in the current velocity and the sedimentation which takes place at these points often results in the formation of a delta. The material transported along the bed is affected very rapidly by variations in the flow conditions and movement of this material ceases as soon as the current velocity falls below the critical limit for transportation, as is often the case at river mouths. On the other hand, the suspended material is affected much more slowly. Since the rate of sedimentation decreases with the particle size, a large proportion of the finer-grained particles especially may be spread out over fairly large areas in front of river mouths.

The extensive delta formations which often arise at river mouths in lakes and seas are therefore an obvious testimony to the eroding, transporting and accumulating action of the rivers and rapidly reflect changes in the activity of the fluviatile, marine and/or lacustrine processes. Continual splitting, formation, oscillation and closing of distributaries characterize the active phase of delta formation, by which the normally branching delta pattern is created and modified.

The modern delta formations, particularly the marine ones, have in recent years been the subjects of increasingly intense study. Many important results have been obtained, but our knowledge of the genesis, morphology and stratigraphy of the different delta types is still far from complete. Opinions differ greatly and are often contradictory, particularly as regards the delta-forming processes. With respect to the morphology of the delta formations, the need of detailed studies would seem to be particularly great as regards the delta front and the areas of bifurcation and re-joining or, in other words, the pattern of branching. Stratigraphic studies have led to a relatively diversified picture of the structural features of many marine deltas (cf. *Shepard*, 1964). Lacustrine deltas, however, have not been subjected to equally detailed study and the interpretation of their stratigraphy given in many geographical and geological manuals would seem to be altogether too schematic and largely misleading.

The majority of delta investigations have been of a morphological and purely descriptive character. In fairly few works have detailed

attempts been made to analyze the relation between the fluviatile processes in the delta-forming distributaries and the appearance and structure of the resulting formations. From the morphological and sedimentological viewpoints, the greatest shortcoming would seem to be that the authors have often not distinguished between the occurrences of bed-transported and suspended material and between the different morphological and structural features which the deposition of these materials usually gives rise to. Our knowledge of the significance of vegetation in the development of the morphological pattern is defective. Similarly the relation between delta formation and climatological environment has been fairly little discussed, principally because our knowledge of delta formation in arctic and tropical regions is still relatively incomplete. For this reason it is at present very difficult to make a detailed classification of different delta types on the basis of studies in the literature.

Besides those in the geographical, geological, oceanographic and hydrological literature, there are numerous papers in the technical and biological literature, that deal with different deltaic problems. Special stability conditions in deltas and deltaic channels that are important to shipping but difficult to navigate have attracted the attention of geotechnicians and hydrotechnicians respectively. Deltas offer soil scientists and biologists many interesting problems. A characteristic of marine deltas, for example, is the gradual changes or rapid fluctuations in the chemical environment, produced by the mixture of fresh and salt water. In this way the conditions for both soil formation and for biological production, colonization and succession are changed.

A review of the earlier literature on deltas and an extensive discussion of the delta-forming processes will be found in *Credner's* monograph *Die Deltas* (1878). A compilation of especially the Russian literature in this sphere will be found in *Samojlov's* informative work on river mouths (1952). Besides the fluviatile and oceanographic processes, *Samojlov* also deals with factors influencing soil formation and biological production. The greater part of his work is devoted to descriptions of the mouths of the larger rivers, in which particularly the mouths of the Russian rivers are treated in detail. Among later Russian papers of more general interest we may mention that of *Egorov* (1955) on general laws for formation of deltaic and littoral plains. A good survey of recent advances in the study of deltaic sedimentation has been written by *van Straaten* (1960). His paper deals especially with marine deltas and with the oceanographic aspects of deltaic sedimentation. The summary is largely based on recent investigations of the deltas of the Mississippi, the Orinoco, the Rhône and the Rhine. Most of the illustrative examples are taken from the Rhône delta, where important investigations have been carried out by, *inter alia*, *Russell* (1942), *Kruit* (1955), *Razavet* (1956) and *van Straaten* (1959 a, 1959 b). Valuable information on both modern and ancient deltaic sediments will be found in the special publication on deltaic and shallow marine deposits edited by *van Straaten* (1964). Geographical, geological, hydrological, pedological and biological characteristics of deltas, especially of the humid tropical zone deltas, were discussed at the Dacca symposium in 1964. The papers presented at this meeting and a summary of the discussion which followed were published in 1966 by *Unesco*.

Of the larger delta formations, that of the Mississippi is probably the best known from the morphological, sedimentological and oceanographic points of view. The literature on this delta has long been particularly extensive, as is clear from the bibliography compiled by *Russell and Dohm* (1936). Amongst the more important morphological works on this area are those of *Russell* (1936) and *Welder* (1959). Important sedimentological and oceanographic investigations have been made by, *inter alia*, *Fisk* (1944), *Fisk, McFarlan, Kolb and Wilbert* (1954), *Shepard* (1956) and *Scruton* (1956). Most of the sedimentological results are summarized in the symposium on recent sediments, northwest Gulf of Mexico, edited by *Shepard, Phleger and van Andel* (1960). More recently important sedimentological works on this area have been published by *Coleman, Gagliano and Webb* (1964) and by *Coleman and Gagliano* (1964, 1965).

Naturally the majority of the large and economically important deltas have been the subjects of very comprehensive investigations. In Fennoscandia there are no modern deltas that are either large or economically important. It is therefore not surprising that the literature on modern deltas from Fennoscandia is relatively meagre. However, the construction of hydro-electric power stations now in progress has in recent years been accompanied by greater prospects (*inter alia*, from the financial point of view) of studying

different deltas, especially with reference to the effect on them of water control. The investigations made and in progress comprise both freshwater and brackish-water deltas and salt-water deltas affected by the tide. A number of these investigations have been published, for example, those of *Rosberg* (1892, 1895), *Ahlmann* (1915), *Arnborg* (1948, 1959), *Säntti* (1954 *a*, *b*), *Axelsson* (1955, 1960, 1963), *Stenborg* (1957) and *Dahlskog* (1963, 1966). On the other hand the literature on ancient glacifluvial deltas is very extensive. Amongst the more important Swedish works are those of *Nelson* (1910), *Hörner* (1927), *Bergsten* (1943) and *Hörnsten* (1964).

The deltas have principally been classified according to the environment, the morphology (for example, the form of the delta front), the slope of the delta plain, the predominant grain size, the rate of advance, the tidal influence, and the stage in the cycle. See, for instance, *Lyell* (1854), *Gulliver* (1899), *Barrell* (1912), *Klenova* (1948), *Samojlov* (1952), *Moravetz* (1960), *Kruit* (1964), *Moore* (1966), *Egorov* (1966) and *Volker* (1966).

From the stratigraphical point of view the deltas can be divided into those with and those without real foreset beds. The pre-requisite condition for a river to be able to form a delta of the classical type, with well-developed foreset beds, as described by *Gilbert* (1885), is that transport along the bed (contact load) should occur right down to the mouth. The occurrence of contact load does not, it is true, always mean an appreciable development of foreset beds but it accentuates and enriches the morphological features of both the delta front and the delta plain and gives rise to more rapid changes in the deltaic river pattern. Deltaic accumulations which are built up of suspended load alone have no real foreset beds and foreset slopes. They might therefore, from both the genetic, the morphological and stratigraphic points of view, be characterized to some extent as incomplete deltas. By foreset beds and foreset slopes I here mean only those parts of the beds or frontal slopes where the dip angles are greater than 10° (cf. p. 24).

The pre-requisite condition for a river to be able to form a subaquatic-subaerial delta has been discussed in detail by *Credner* (1878). In this connection he came to the conclusion that only local significance can be assigned to the sediment transport of rivers, their current velocities, the depth conditions in front of the mouths and the mechanical action of the sea in its influence on the delta formation. On the other hand, he considered that the changes in level were of decisive importance for the delta formation and for the geographical distribution of the deltas. On coasts of emergence deltas may be formed even under otherwise unfavourable conditions. On coasts of submergence, on the other hand, delta formation is often precluded.

These conclusions of *Credner's*, which were based on, *inter alia*, collations of partly erroneous data on coasts of emergence and submergence, have been the subject of lively discussion and criticism (cf., for instance, *Gulliver* 1899 and *Samojlov* 1952). It is clear that these conclusions cannot apply generally from the fact that rivers which carry large amounts of sediment may form extensive deltas even on coasts of submergence. One of the foremost examples of this is the Mississippi delta (*Russell*, 1936).

By applying the theory of jet flow to delta formation, *Bates* (1953) has been able to analyse more thoroughly and from new points of departure the processes which take place at the mouths of the delta-forming distributaries and in this way has attempted to explain the genesis of the different delta types which in his opinion exist. *Bates's* paper gave rise to a discussion between *Crickmay and Bates* (1955), in which *Crickmay* sharply criticized the way in which *Bates* had applied the theory of jet flow to the delta-forming processes. *Scruton* (1956) was also very critical of *Bates's* theory of delta formation and questioned the correctness of *Bates's* conclusions.

Though *Bates* obtained some strange results through not taking proper account of the morphology of the river mouths and the processes of transportation and deposition of bed load and suspended load, his approach to the problem is probably a good way of reaching a better understanding of the mechanics of delta formation. The expanding jet flow over a delta front and the phenomenon of flow separation have been treated more in detail by *Jopling* (1960) in his fundamental treatise on the mechanics of bedding.

The formation and development of distributaries, crevasses and subdeltas are intimately related to the changes in the slope of the energy gradient which are produced, for example, by the elongation of the distributaries and by the variations in the water level. Especially the transportation of bed load varies rapidly with changes in the bed tractive stress which thereby arise, and even

a slight damming, produced, for example, by the effect of wind, may lead to a rapid blocking and closing of the distributaries which debouch in the actual or prevailing wind direction. This is one of the more important reasons why the deltaic pattern is usually changed more rapidly in the parts in which bed-load transport occurs than in the parts in which there is only suspended-load transport.

The morphological pattern at the river mouths is largely formed by conjoint action between the fluviatile and the marine or lacustrine processes. The more powerful the shore processes are in relation to the fluviatile processes, the more obvious will be the general features of the shore morphology. When the shore processes are so predominant that the river can only be regarded as a supplier of the material taken over by the waves and tidal currents, we can no longer speak of a delta.

In studying deltaic formation, the fluviatile processes should therefore be regarded as primary. In what follows, the morphological delta pattern will, for this reason, be considered principally from fluviatile points of view.

Valuable information on both modern and ancient deltas will also be found in the special publication on deltas in their geological framework recently published by *Houston Geological Society* (1966).

Chapter 2

THE DELTA FRONT

Flow separation and jet flow

The rapidly changing boundary conditions at the mouths of distributaries is accompanied by a pronounced modification in the flow pattern. The flow separates from the diverging boundaries of the distributary mouths and the flow pattern over the delta front simulates that of expanding jets with complementary zones of reverse flow, characterized by appreciable circulation and large-scale eddying. These hydraulic phenomena in the immediate vicinity of the distributary mouths play an important part in the processes of transportation and deposition of bed load and suspended load and are therefore important for the morphology and stratigraphy of deltas.

The flow pattern at the delta front is very complicated and a theoretical solution to the problem of velocity distribution and turbulence structure in the flow regions at and beyond the distributary mouths is not yet available. The flow from the distributaries, however, bears a similarity to the expansion of submerged jets and valuable information on the flow conditions may therefore be given by theoretical and experimental studies of submerged jets.

Experimental investigations of the flow in convergent and divergent channels have been carried out by, among others, *Nikuradse* (1929). His results indicate that separation may occur when the half-angle of divergence is about 5 degrees. The conditions of flow separation are also influenced by the state of flow and by the configuration of the diverging boundaries. In actual practice, however, a divergent boundary, at an angle exceeding 10 degrees, normally fails to constrain the flow, and separation results (*Jopling*, 1960).

If the separated jet flow from a circular orifice can spread out in all directions, it forms a three-dimensional, symmetrical axial jet. If the expansion of a jet from an elongated slot orifice is only two-dimensional, a plane jet is formed.

The fluid dynamics of jet expansion were first investigated theoretically by *Tollmien* (1926). He applied *Prandtl's* (1925) mixing-length theory to solve the mixing problems of a jet issuing into a medium at rest. *Tollmien's* results were extended to the case of a jet issuing into a medium not at rest by *Kuethe* (1935) and by *Squire and Trouncer* (1944). *Tollmien's* theory has been further developed and tested experimentally. Detailed discussion and relevant references may, for instance, be found in *Förthmann* (1934), *Reichardt* (1941, 1942), *Albertson, Dai, Jensen and Rouse* (1950), *Elrod* (1954), *Pai* (1954), *Wiegel* (1964) and especially for the delta case in *Bates* (1953) and *Jopling* (1960).

The flow pattern of a jet can be subdivided into two zones. The first of these zones is called the zone of flow establishment and the second the zone of established flow. Fig. 1, modified after *Albertson et al.* (1950), illustrates the pattern of diffusion of a submerged jet issuing into a medium at rest. Beyond the efflux section the pronounced velocity discontinuity between the jet and the surrounding fluid is manifested in the form of eddies of turbulent mixing generated at the borders of the jet. As the direct result of this the fluid within the jet will undergo both lateral diffusion and deceleration. Surrounding fluid is not only accelerated but also entrained and therefore the volume of flow and the breadth of the jet increase with the distance from the efflux section.

The initial zone of flow establishment is characterized by an inner core, where the longitudinal velocity (u_x) remains constant, being equal to the efflux velocity (u_o). This inner constant-velocity core decreases in lateral extent and the terminal point is reached when the eddies of lateral mixing have penetrated to the centre-line of the jet. Thereafter the diffusion process continues without essential change in character and the flow may be considered as fully established.

According to *Albertson et al.* (1950), the ratio of the length of the zone of flow establishment to the depth or diameter of the efflux section (x_o/D_o) is 5.2 for the two-dimensional jet and 6.2 for the three-dimensional jet, which implies that the boundary of the constant-velocity core is tilted at an angle of 5.5 and 4.6 degrees respectively to the centre-line. For the zone of established flow

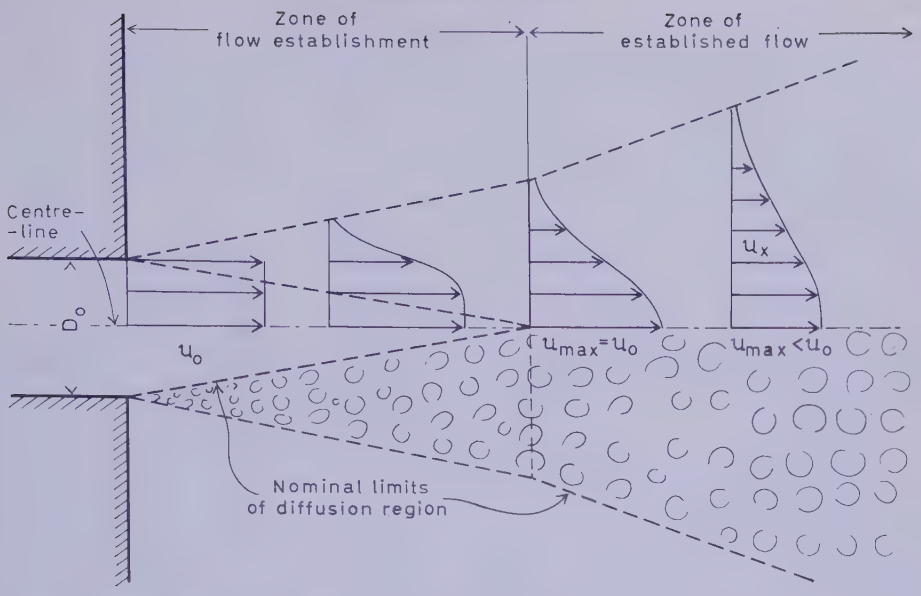


Fig. 1. Schematic representation of jet diffusion. (Modified after *Albertson et al.*, 1950.)

the nominal boundary of the two-dimensional jet has a slope of approximately 1: 4 (14 degrees), whereas that for a three-dimensional jet is approximately 1: 5 (11—12 degrees).

It must, however, be pointed out that the border between the zones cannot be sharply defined, because a zone of transition necessarily exists between two distinct flow regimes. "There is, in fact no precise point at which the eddies from opposite sides of the jet can be said to meet, for the statistical nature of the mixing process makes it impossible to place more than an arbitrary limit on its lateral range. The boundaries of the diffusion region, like the border between the two zones, must therefore be accepted merely as convenient nominal disignations" (*Albertson et al.*, p. 641).

The measurements of the centre-line velocity (u_{max}) in the zone of established flow by *Albertson et al.* gave the following equation for the three-dimensional jet (circular aperture)

$$\frac{u_{max}}{u_o} \cdot \frac{x}{D_o} = 6.2 \qquad (1)$$

and the following for the two-dimensional jet (infinite slot)

$$\frac{u_{max}}{u_o} \sqrt{\frac{x}{D_o}} = 2.28 \qquad (2)$$

Elrod (1954) based his computations of jet velocities and entrainment ratios on *Reichardt's* (1941) free turbulence theory. According to him, the longitudinal velocity (u_x) at a point with coordinates x, y, z in jet flows from rectangular apertures of dimensions $D_o \cdot b_o$ (depth · width) may be expressed by the equation

$$\frac{u_x}{u_o} = \tfrac{1}{2} \sqrt{\left[\Phi\left(\frac{y+\frac{1}{2}D_o}{Cx}\right) - \Phi\left(\frac{y-\frac{1}{2}D_o}{Cx}\right)\right]} \cdot$$

$$\cdot \sqrt{\left[\Phi\left(\frac{z+\frac{1}{2}b_o}{Cx}\right) - \Phi\left(\frac{z-\frac{1}{2}b_o}{Cx}\right)\right]} \qquad (3)$$

For the centre-line velocity equation (3) may be written

$$\frac{u_{max}}{u_o} = \sqrt{\Phi\left(\frac{D_o}{2Cx}\right)\Phi\left(\frac{b_o}{2Cx}\right)} \qquad (4)$$

In equations (3) and (4) the Φ function is the probability or error integral,

$$\Phi(\gamma) = \frac{2}{\sqrt{\pi}} \int_o^{\gamma} e^{-t^2} dt$$

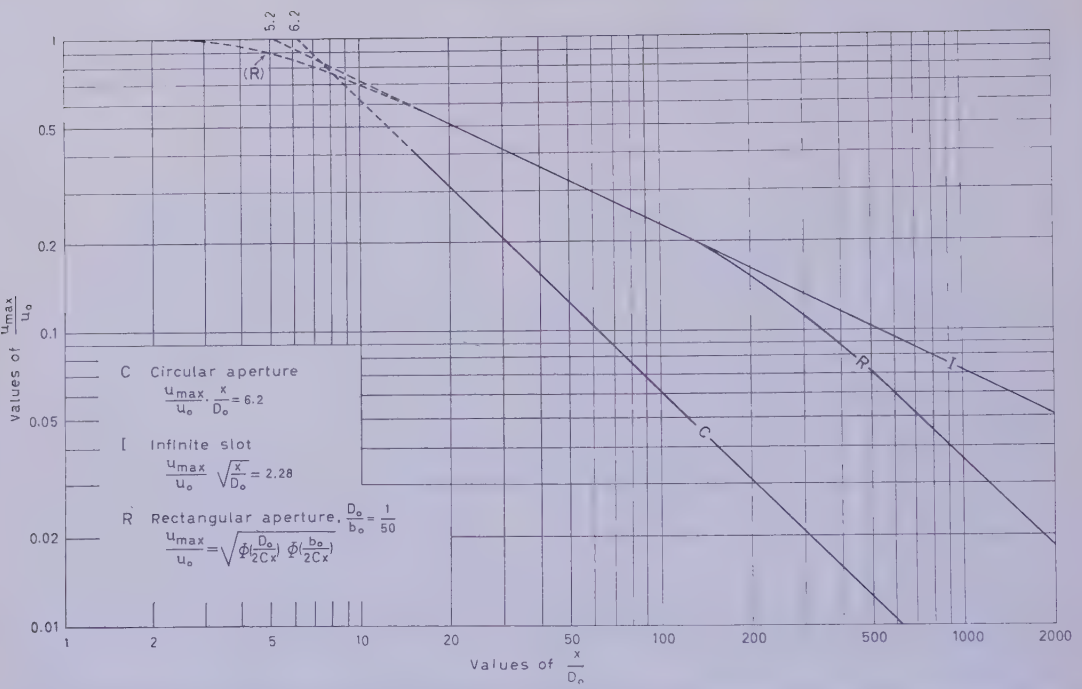


Fig. 2. Distribution of centre-line velocities for flow from circular aperture, infinite slot and from rectangular aperture with $D_o = b_o / 50$.

and is tabulated in many statistical handbooks. See, for instance, *Czuber* (1914). For the empirical constant C in equation (4) the value of 0.109 is used here, that is, the same value as *Albertson et al.* (1950) found for the two-dimensional jet.

In Fig. 2 the centre-line velocity for the rectangular jet has been computed for an aperture with $D_o = b_o / 50$. As may be seen from the figure, the zone of established flow for a rectangular jet may be subdivided into a zone of partly established flow and a zone of fully established flow with an intervening second zone of transition. In the zone of partly established flow, the length of which depends on the geometry of the aperture, the sloping line for the centre-line velocity follows the line for a two-dimensional jet, which, apart from the zone of transition, has a slope of 1: 2. In the zone of fully established flow the sloping line for the centre-line velocity goes parallel to the line for a three-dimensional jet, which has a slope of 1: 1.

Bates (1953) utilized the jet theory in his discussion of delta formation. The basic types of inflow defined by *Bates and Freeman* (1953) were described by *Bates* (1953, p. 2125) as follows:

"*a*. Hyperpycnal inflow (inflow more dense), where the sediment-laden fluid flows down the side of the basin and then along the bottom as a turbidity (density) current. With vertical mixing inhibited during this flow, because the dense fluid seeks to remain at the lowest possible level, the flow pattern is that of the plane jet. The most frequent site for delta formation by such flow is at the mouth of submarine canyons.

b. Homopycnal inflow (inflow equally dense), where sediment-laden fluid enters a basin filled with fluid of comparable density, as in the case of a stream entering a fresh-water lake. Mixing takes place readily in three dimensions, and the flow pattern is that of the axial jet. The type of delta which forms is the classical type with top-, fore-, and bottom-set beds as described by *Gilbert* (1885).

c. Hypopycnal inflow (inflow less dense), where sediment-laden fluid moves out over the surface of denser

> fluid filling the basin, as in the case of a stream discharging into the ocean. Vertical mixing is inhibited because of stability between the layers, and the flow pattern is that of the plane jet. If the magnitude of discharge is small, a lunate bar forms off the outlet; if the discharge is moderate to large, a cuspate, or bird-foot type delta will form."

Starting from *Leighly's* (1934) principle that stream deposits are formed in areas lateral to threads of maximum velocity and its flanking threads of maximum turbulence, *Bates* concluded that submerged natural levee formations would be highly developed along the flanks of the jet to a distance of 4 channel depths and 4 channel widths out from the orifice when the flow pattern was that of the axial (three-dimensional) jet and that of the plane (two-dimensional) jet respectively. He further concluded that there would be no deposition in the constant-velocity core zone but that in the transition zone 4—8 channel depths and 4—8 channel widths out from the orifice in case of axial and plane jet flow respectively, deposition of bed-load and suspended load would take place, because deceleration of the flow along the centre-line of the jet now begins. This deposition creates a lunate bar, blocking the mouth of the outlet and causing the channel to split into two or more distributaries.

Because the depth of the distributary mouths is usually only a small fraction of the width, *Bates* suggested that the rate of deposition close to the distributary mouths and the frequency of channel splitting would be much higher if the flow were that of the axial jet than if the flow were that of the plane jet. He therefore concluded (p. 2132)

> "that homopycnal inflow develops partly subaerial, flat topset beds flanking the stream channels, and that these beds change quickly to fore-set beds resting at the angle of repose immediately outside the distributaries' mouths. Likewise, there is limited deposition of fine-grained material in the deep part of the basin as thin bottom-set beds. Over a period of time, the continually shifting distributary outlets also build an arcuate, symmetrical deltaic deposit off the initial outlet of the master stream. In other words, the depositional pattern for homopycnal inflow agrees closely with that reported by *Gilbert* (1885) as occurring in nature."

Scheidegger (1961, pp. 249—250) accepted *Bates's* theory of delta formation but concluded that axial jet flow leads to the formation of arcuate deltas and that bird-foot deltas are the result of plane jet flow. This conclusion is, however, not correct, as will be discussed later.

In the discussion between *Crickmay and Bates* (1955), *Crickmay* suggested that jet diffusion will have a plane or two-dimensional pattern, whether the density of the inflow is the same as that of the basin or not. He further suggested that, for slot-like river mouths, depth is the only dimensional factor that has a bearing on velocity and carry of jet flow. He also pointed out that natural inflows are vastly more complicated than free jets on account of their boundary geometry and that diffusion patterns and manner of flow become complicated accordingly. *Bates* admitted that the geometry of the river mouths affects the pattern of diffusion but considered that his use of the jet theory was in principle correct.

The theory of jet flow gives no doubt valuable information on the flow pattern at the delta front and therefore also on the pattern of deposition. It must, however, be emphasized that it is also necessary to pay attention to several other factors, such as, for instance, the geometry of the river mouths, the depth conditions, the frontal shape of the delta, the grain size of the transported and deposited material, density differences, macro-turbulence, waves and water-level variations.

The distributary-mouth bar is, as *Scruton* (1956, p. 2917) points out, part of the distributary orifice. Therefore it cannot be the result of jet characteristics in the way that *Bates* suggested. That *Bates's* conclusions regarding the pattern of deposition are not correct is clear from the fact that deltas of the classical type with real foreset beds and deltas without real foreset beds are built up both by rivers with a homopycnal type of inflow and by rivers with a hypopycnal type of inflow. An example of a fresh-water delta without real foreset beds but with the homopycnal type of inflow is the delta of the river Kemijoki in Lake Kemijärvi in northern Finland (*Axelsson*, 1964). As an example of a salt-water delta with real foreset beds but with the hypopycnal type of inflow, we may mention the delta of the river Rombakselva on the coast of northern Norway. The steeply inclined (up to 30 degrees) foreset

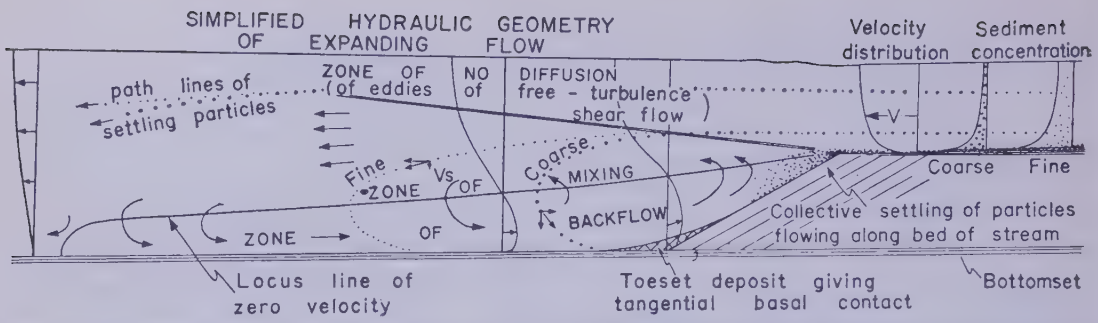


Fig. 3. Zonal terminology for rapidly varying flow over the foreset slope of a small delta. Dotted lines represent idealized path lines of settling particles. (From *Jopling*, 1963.)

slopes and beds of this delta have a maximum height of more than 15 metres (*Marklund*, 1960).

The rate of deposition close to the distributary mouths and the frequency of channel splitting is not only dependent on the type of inflow but also on many other factors, of which the grain-size distribution of the transported material is probably one of the most important. For a river to be able to form a delta with real foreset beds, the most important condition is not, as *Bates* suggests, that the inflow should be of homopycnal type, but that there should be bed-load transportation right up to the mouth.

As may be seen from Fig. 2, the distribution of centre-line velocity in the zone of partly established flow of a rectangular jet follows that for a plane (two-dimensional) jet with the same slot dimension as the depth of the rectangular aperture. It is also evident that the length of the inner constant-velocity core and the deceleration of centre-line velocity in the zone of partly established flow is a function of the depth of the rectangular aperture.

Distributary mouths are usually more or less rectangular, the depth being small in relation to the width. On the other hand, the depth ratio and the dip angles in front of the distributaries vary a great deal. Coarse-grained deltas are usually characterized by low depth ratio and high dip angles, fine-grained deltas by high depth ratio and low dip angles and the discontinuity in the frontal profile may be entirely absent. The term "depth ratio", used by *Jopling* (1960, 1963), here means the ratio of distributary mouth depth to basin depth in front of the distributaries, as, for instance, the depth of the delta toe.

When the depth ratio is low and the dip angles high, the abrupt vertical boundary expansion causes the flow to separate from the bottom and the flow pattern over the foreset slope bears a similarity to that of a rectangular jet. The angle of tilt of the diffusion boundary to the surface and the deceleration of centre-line velocity close to the distributary mouth is now influenced by the depth of the distributary mouth.

When the depth ratio is high and the dip angles low, the flow separation is only lateral and the flow pattern is therefore similar to that of a plane (two-dimensional) jet. The deceleration of centre-line velocity is slower, now being influenced by, *inter alia*, the width of the distributary mouth.

For the homopycnal type of inflow the flow pattern should not be compared with that of a submerged axial jet, as *Bates* (1953) suggested, but with that of a rectangular jet or with that of a plane jet, depending on whether there is an abrupt vertical boundary expansion or not.

One also has to consider that flow expansion in a lateral direction often begins upstream of the foreset slope, and that flow expansion in a vertical direction is restricted beyond the foreset slope. The expansion may therefore be two-dimensional upstream of the delta lip, three-dimensional over the foreset slope and two-dimensional beyond the foreset slope.

For hyperpycnal and hypopycnal types of inflow the similarity with submerged jet diffusion is less. Even small density differences affect the flow pattern and the mixing process and, as *Sundborg* (1956, p. 164) points out, "the way in which a density current spreads out is quite different from that of a free jet".

Elrod (1954, p. 151) states that "a jet may be considered as free until the cross-sectional area

of the jet equals about one-fifth of the total cross-sectional flow-area of the region through which it is proceeding". The limited nature of the expansion in front of the distributaries therefore implies that the flow may not be considered as a free jet.

Several of the fundamental differences that distinguish the delta from the jet case are discussed by *Jopling* (1960), who, from a two-dimensional approach to the problem, carried out experimental observations on the flow separation over a deltaic channel discontinuity.

For deltas with low depth ratio and high dip angles the free surface status implies that the zone of no diffusion is about twice as long as that for a rectangular jet from an aperture of the same dimensions as the half-jet of nature. *Jopling's* experimental results demonstrated that the diffusion boundary, representing the limit of penetration of the eddies of free-turbulence shear (see Fig. 3), is tilted at an angle of approximately 5.5 degrees to the horizontal. The value is in close agreement with that for a free jet. The velocity in the zone of no diffusion was, however, subject to a decrease away from the delta lip, accompanied by a compensatory rise in water level. The magnitude was dependent, *inter alia*, on the Froude and Reynolds numbers of the flow, but for subcritical transitions (the norm in nature) the surface disturbance and the surface and subsurface decrease in velocity were slight within the zone of no diffusion. For relatively high depth ratios, however, the terminal part of the zone of no diffusion was obliterated by the eddies of mixed flow.

In the restricted delta case the solid boundaries interfere with the entrainment process and demand the establishment of return flows. In *Jopling's* experimental investigations the average maximum velocity of backflow, calculated from vertical velocity profiles, was approximately one-fifth to one-quarter of that of the average velocity upstream of the delta lip. The velocity of the fastest filament in the zone of backflow, enclosed between the main stream and the bottom, was, however, variable. Rapid acceleration occurred at the downstream extremity and deceleration up the frontal slope of the delta. The intensity of turbulence increased up-slope, whereas the scale of turbulence decreased. Bed-load was dumped on the upper foresets. Some of the suspended particles by-passed the zone of backflow and some was deposited within the zone of backflow. The approximate path lines of suspended particle movement were traced by plotting the vectorial resultant of the longitudinal velocity of the stream and the nominal settling velocity (v_s) of the particle. Increasing flow velocity, suspended load concentration and depth ratio favoured the deposition in the toe sector and the development of a tangential basal contact. See Fig. 3, from *Jopling* (1963).

In natural river mouths the flow velocity, the sediment concentration and grain-size distribution vary both in the vertical and lateral directions and with time. The bed configuration varies and is influenced by the migration of bars that may deflect the flow direction. The lateral zones of flow separation and backflow vary with the variation in water level. Wind, waves and basin currents influence the flow velocity and the flow direction in front of the river mouths. Density differences, sediment concentration, temperature and salinity affect the flow pattern, the mixing process and the rate of deposition. Ice, vegetation and irregularities of the basin bottom also affect the flow etc.

Fine-grained deltas are generally characterized by high depth ratios and gently inclined frontal profiles. Flow separation is now only lateral and the flow over the delta front is therefore more gradually varied and associated with a backwater effect, which causes a slow progressive reduction in flow velocity and bed tractive stress in the downstream direction and therefore also influences the deposition and sorting of the transported sediment. The limit of this backwater effect coincides with the theoretical upstream limit of topset aggradation and with delta advance it retreats further upstream (*Lane*, 1955).

The flow pattern and the pattern of deposition in front of the distributaries may thus be very complicated in detail. For the homopycnal type of inflow, however, the jet analogy no doubt gives valuable information on the general flow conditions in the immediate vicinity of the distributary mouths and therefore also for the general trends of deposition and sorting of suspended sediment. "The marginal diffusion and deceleration of the expanding flow results in a reduction of flow competency and capacity" (*Jopling*, 1963, p. 118), and this is especially the case in deltas with distributary mouth bars or greater distal lobes of bed-load origin and with rapidly varying flow over the delta front.

Density differences and density currents

The density of water and the density difference between adjoining water layers affect the turbulent velocity fluctuations and the settling velocities for sediment particles. The turbulent exchange decreases with increased density stratification and density differences occurring between inflowing river water and adjoining basin water therefore restrict the flow expansion and influence the rate of deposition in front of the distributaries and the morphology of the delta front. If the density contrast between contiguous bodies of water is large, the turbulent exchange will usually be insignificant and so-called density currents will arise. A discussion of the effect of density stratification in damping turbulent exchange and relevant references will be found in *Sundborg* (1956), in *Harleman* (1961) and in several papers presented at the 8th and 11th congress of the *International Association for Hydraulic Research* (1959, 1965). Valuable information on density currents, especially the type known as turbidity currents, is summarized in the symposium on turbidity currents edited by *Hough* (1951), in the comprehensive survey of sedimentation in Lake Mead by *Smith, Vetter, Cummings, and others* (1960), and in the special publication on turbidites edited by *Bouma and Brouwer* (1964) which also contains a comprehensive bibliography of turbidity currents and turbidites compiled by *Kuenen and Humbert* (1964).

The density of water varies with the temperature, the content of dissolved and suspended material and with the pressure. The effect of differences in temperature, dissolved salts and suspended sediment upon the density contrast between adjoining water layers is illustrated in Fig. 4. The effect of temperature upon the density of pure water is tabulated by *Welch* (1948, p. 350). The values for dissolved salts refer to those for sea water (see *Knudsen*, 1901) and they may therefore differ somewhat from those for river and lake water, owing to the different ionic composition. No account was taken of the fact that the relation between the content of dissolved salts and the density differs somewhat from a linear function and therefore the values given should only be regarded as approximate. The values for suspended sediment are calculated for material of a density of 2.7 g cm^{-3}.

As may be seen from Fig. 4, one and the same difference in temperature causes a greater density difference at high than at low temperatures. The density difference caused by a temperature difference of 1 degree, for instance, is about three times greater at 25 degrees than at 10 degrees. At high temperatures, therefore, even small temperature differences may cause considerable damping of the turbulent exchange.

Density stratification in rivers is generally caused by suspended sediment, in lakes and in the sea by temperature and salinity. At river mouths the density contrast between the inflowing river water and the entraining basin water may periodically be considerable. The flow from deltaic distributary mouths may therefore be characterized by density currents, and this is normally the case as regards the distributaries that discharge their waters into the sea.

In Lake Laitaure in northern Sweden a density contrast as low as 0.0003 between the inflowing river water and the impounded lake water is sometimes enough for the formation of density currents (see further p. 112). As may be seen from Fig. 4, a difference in suspended sediment of 500 g/m^3, in dissolved salts of 400 g/m^3, or in temperature of 1—2 degrees at high water temperatures may thus be enough for the formation of density currents. Even in several fresh-water lakes, density currents should therefore sometimes exist.

According to a definition by *Bell* (1942 *a*, p. 514), "a density current may be described as a gravity flow of a liquid or a gas through, under, or over a fluid of approximately equal density". On the basis of the position of the density currents relative to the boundaries of the surrounding fluid we distinguish between overflows (surface flows), interflows and underflows (bottom flows). An underflow is usually implied if the position is not indicated. The density currents that have a higher content of suspended sediment than the enclosing fluid are called turbidity currents or suspension currents. These currents also differ from other density currents, because the density is not uniform but increases from top to bottom (cf. *Bell, 1942 b; Menard and Ludwick*, 1951; *Gould*, 1951).

From the morphological and sedimentological points of view, density currents of the turbidity underflow type are usually paid most attention. In Fig. 5, modified after *Knapp* (1943), the beginning of a turbidity current underflow is illustrated. The turbid flow leaves the surface at

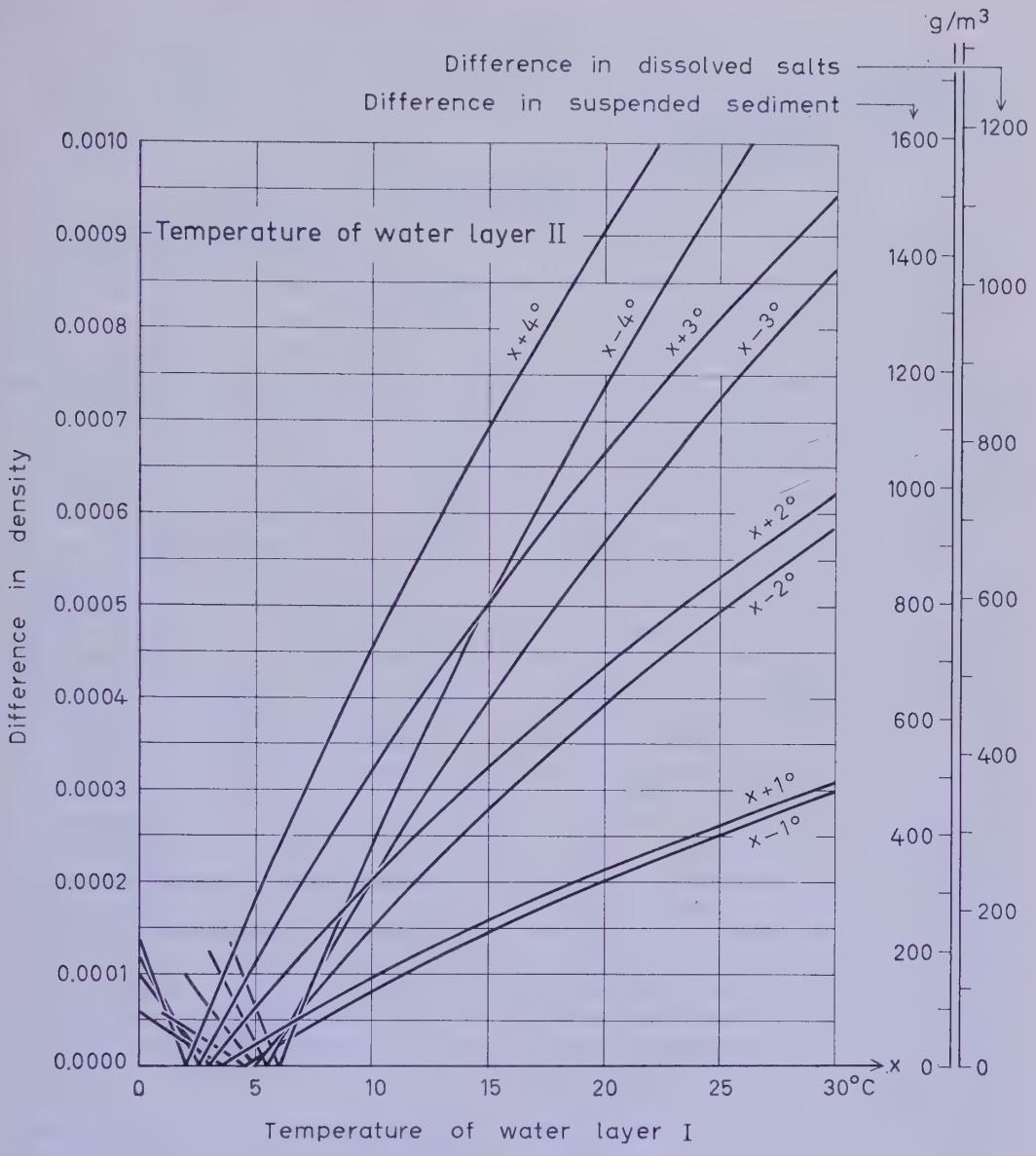


Fig. 4. Effect of differences in temperature, dissolved salts and suspended sediment upon density differences between adjoining water layers.

the plunge point and continues as an underflow in the reservoir or basin. On account of the shear at the interface between the underflow and the overlying basin water, some of the basin water is dragged down and a corresponding counterflow is induced in the upper portion of the basin. In nature the plunge point is therefore often marked by floating debris, since surface circulation flows towards this point and then dives below.

According to *Daly* (1936) and *Kuenen* (1951, 1952), the velocity of a turbidity underflow may be approximated by a Chézy-type formula

$$u = C \sqrt{R_d \operatorname{sine} S d_e} \qquad (5)$$

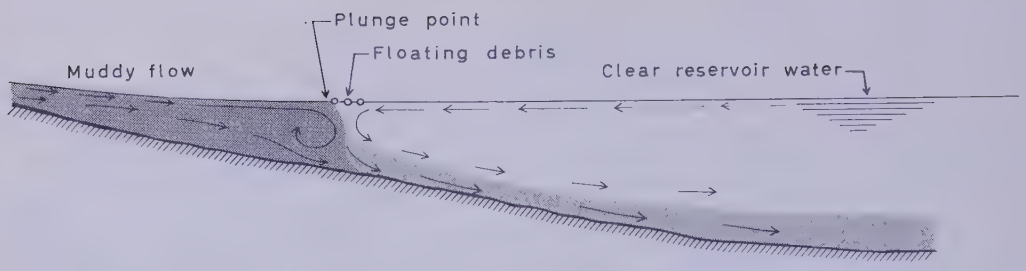


Fig. 5. Section of reservoir showing beginning of underflow. (Modified after *Knapp*, 1943.)

where R_d is the hydraulic radius of the density current, S the slope angle, d_e the effective density (density contrast between the current and surrounding fluid), and C an empirically determined constant. The conventional hydraulic radius (R) is obtained by dividing the cross-sectional area (A) by the wetted perimeter (P). For a density current, however, a hydraulic radius ordinarily means a special one (R_d), in which the cross-sectional area is divided by that portion of the perimeter which is in contact with other fluid masses (cf. *Bell*, 1947). According to the formula, the flow velocity of a turbidity underflow thus increases with the size of the current, the slope, and the effective density. This relation has been treated theoretically by *Hinze* (1960) and by *Plapp and Mitchell* (1960).

The persistence of a density current will depend on the energy gradient and the effective density. The density of a turbidity current may be lessened through deposition of some of the suspended sediment. Especially coarse-grained turbidity currents therefore are very sensitive to changes in the state of flow. The most common cause of lessening or eliminating of the density contrast is mixing with surrounding fluid. The degree of mixing therefore is of vital importance for the formation and the persistence of density currents.

The mixing characteristics of density currents were outlined and discussed by *Knapp* (1943). He distinguished between localized mixing and general mixing and subdivided the former into entrance mixing, plunge-point mixing, and mixing due to channel irregularities.

The general mixing across the boundary or interface between the density current and the surrounding fluid "is a wave phenomenon and is governed by the normal laws of surface waves applied to the physical conditions of density flows" (*Knapp*, 1943, p. 299). If the difference in the velocities of the two fluids is greater than the velocity of these waves, mixing of the two fluids commences.

The combined effect of entrance and plunge-point mixing on the size of density currents in Shaver Lake was studied by *Bell* (1947). In this lake the density currents are of a thermal nature and of interflow type. On account of entrance and plunge-point mixing the volume of flow increased and the size of the thermal density currents were from less than twice to nearly five times the size of the inflowing streams. The density difference between the density current and the unmixed lake water was sometimes as low as 0.0001.

According to *Bell* (1947), the hydraulic radius, the effective density and the available specific energy are the chief factors influencing entrance and plunge-point mixing. Any increase in the effective density or in the hydraulic radius tends to decrease mixing, whereas any increase in specific energy tends to increase it. Small inflows tend to mix more readily than large ones, because the area of contact per unit volume increases as the size of the flow decreases.

Both the hydraulic radius and the flow velocity normally increase with increasing discharge. In Shaver Lake the mixing ratio decreased consistently as the discharge increased and the effect of increasing specific energy on the mixing ratio was therefore less than the effect of increasing volume of flow. The critical value of the effective density for producing density currents may therefore be lower at high than at low river discharges.

The measurements in Lake Laitaure indicate that the morphology in front of the distributaries, such as, for instance, the depth ratio, also influences the mixing ratio. The mixing ratio is higher in front of the distributaries where the

foreset slopes are high than in front of those where the foreset slopes are low or absent. The observations also showed that, especially at high water discharges and at places with a high depth ratio, the plunge-point was situated at a fairly great distance in front of the distributaries, which implies that the lake water was flushed out or greatly diluted in front of the distributary mouths (see further p. 112).

The flow close to deltaic distributary mouths may therefore sometimes be of homopycnal type, even if the density contrast between inflowing river water and entraining basin water is enough to produce density currents and the flow at some distance from the distributary mouths is of underflow (hyperpycnal) or of overflow (hypopycnal) type.

Often, however, relatively unmixed basin water is present close to the distributary mouths and it may also be present in the lower parts of the distributaries when the density currents are of overflow type. On marine deltas, saline basin water buoys up the river water and flows into the river channels, where it may extend upstream for many miles when river discharges are low. These salt wedges may also oscillate upstream and downstream with tidal changes (cf. *Scruton and Moore*, 1953; and *Scruton*, 1956).

Where salt wedges are present, the bottom velocities are low and the transportation of bed load therefore usually stops. Some of the colloidal suspended sediment may be flocculated and settle out rapidly. This recently flocculated material, termed ooze, is more easily moved than sediment that has recently settled without flocculation. According to *Bates* (1953, p. 2143), it is therefore easily flushed from the channels when river discharges are rising. Part of this flushed material may probably continue seaward as a turbidity current.

In fresh-water lakes the inflowing river water often has a greater density than the lake water and underflows may be produced. The deceleration of the bottom velocities in front of the distributary mouths is then slower than when the inflow is of homopycnal or of overflow type. With increasing bottom velocities (bed shear) the angle of the foreset slopes systematically decreases. The angle of the foreset slopes may therefore be lessened when the inflowing river water has a higher density than the basin water.

The mixing ratio and the expansion of the current is decreased when the density contrast is increased. The rate of deposition close to the distributary mouths is slower when the suspended sediment is transported by overflow currents than when the inflow is of homopycnal type. Turbidity underflows reduce the dip angles of foreset slopes and favour the development of a strongly tangential contact with the bottomset slopes (cf. *Jopling*, 1960).

Turbidity currents may also be produced in front of the distributaries by settling of sediment from turbid river water, by waves stirring up the bottom sediments, by earthquake shocks, and by subaquatic mass movements in the form of slides, slumps and mud flows along the delta-front slopes (cf. *Menard and Ludwick*, 1951). *Shepard* (1955, 1962) suggests that the delta-front gullies bordering parts of the Mississippi delta are due to mass movements probably of the earthflow type, while *van Straaten* (1959) explains most of the gullies off the Rhône delta as the result of turbidity currents.

Density currents may thus have a great effect on the frontal profile development of a delta but for a better understanding of their importance further experimental investigations and field observations are necessary.

Frontal shape

The form of the delta front is dependent on the combined action of fluvial processes and shore processes, on the ratio between transportation and deposition of bed load and suspended load and on the original topography of the basin.

The fluvial processes are the cause of lobate deposits in front of the distributaries. Through shifting or splitting of these distributaries a broad fan-shaped, multilobate deposit may be formed. The shore processes tend to smooth out any irregularities and the shorelines between the distributaries may therefore be rather straight or cuspate.

On small coarse-grained deltas the foreset slopes may sometimes be rather continuous. The frequency of channel splitting is often high and the shorelines may be rather regular. On large fine-grained deltas the foreset slopes usually occur only at the active distributary mouths or they may be entirely absent. The frequency of channel splitting is often low, the interdistributary bays are usually shallow and the shorelines may be rather irregular.

With increasing wave action, normal beach

accumulations will be more common and longshore transportation may be rather high where sufficient sand is available. Distributary mouths may be curved in the direction of the predominant littoral drift. Spits and bars may be formed between the distributary mouths, thus straightening the shoreline, and behind them lagoons will be enclosed. Where the energy of the waves is great and the bottom slope is low, several successive beach ridges, separated from each other by swales, may characterize large parts of the frontal delta platform.

Where the tidal range is great, the distributary mouths are usually widened and have estuarine properties. The beaches between the estuarine distributary mouths are often relatively short and separated from each other by tidal inlets, which give access to tidal flat areas.

The original inclination of the basin floor usually influences the inclination of the bottomset slopes, and especially in shallow basins the form of the delta front may be rather complicated due to initial topographic irregularities.

Foreset slopes and bottomset slopes

The frontal profiles or slopes of a delta are usually subdivided into foreset slopes and bottomset slopes. The terms "foreset slopes" and "bottomset slopes", as well as the related terms "foreset beds" and "bottomset beds", however, have been used rather differently by different authors.

By "foreset slopes of a fresh-water delta" are often meant those parts of the delta front outside the distributary mouths which incline at a high angle, and by "bottomset slopes" the gently inclined slopes beyond the foreset slopes. Often it is suggested that foreset slopes and foreset beds, resting at the angle of repose, are characteristic of fresh-water deltas (cf. *Bates*, 1953).

The angle of discordance between different parts of the frontal profiles of a large fine-grained marine delta may be very slight. By "foreset slopes of a marine delta" are therefore often meant very gently inclined slopes, characterized by a great homogeneity of predominantly fine clastic sediments and by a low content of marine fossils. By "bottomset slopes" are meant the slopes beyond the foreset slopes. They are also characterized by predominantly fine massive sediments but they differ from the foreset slopes by having a high content of marine fossils. The inclination of both the foreset slopes and the bottomset slopes of a marine delta may therefore be very gentle and is often much lower than 1 degree. Sometimes the platforms directly outside the distributaries, characterized by laminated zones with alternating finer and coarser sediments, and the marsh areas, characterized by structureless deposits, are considered to belong to the topset slopes and topset beds of a marine delta (cf. *Shepard*, 1956, 1960; *van Straaten*, 1960; *Scruton*, 1960).

It must, however, be pointed out that the angle of inclination of the frontal slopes is also rather low for several fresh-water deltas and rather high for several marine deltas (cf. p. 17). From the sedimentological point of view it may sometimes be useful to define foreset slopes and bottomset slopes, as mentioned above, for large fine-grained marine deltas. However, from the morphological point of view, as in the present paper, it would be better to use the term "foreset slopes" only for those parts of the delta front, which incline at a relatively high angle, and the term "bottomset slopes" for gently inclined slopes beyond the foreset slopes.

If the frontal slope over which the flow is expanding inclines at an angle steeper than 10 degrees, separation normally results and a zone of backflow is established in the lee of the discontinuity (cf. p. 14). *Jopling* (1960) states that the dip of a bed-load transport slope only exceptionally exceeds 10 degrees and that the majority of occurrences of cross-bedding develop in response to the phenomenon of expanding flow and separation. He therefore takes the phenomenon of flow separation as a genetic distinction between cross-bedding (e.g. foreset beds) and horizontal bedding (e.g. bottomset beds).

This seems to be basically sound and here therefore, for the sake of simplicity, I mean by "foreset slopes" and "foreset beds" only those parts of the frontal slopes and beds of aggradational origin where the dip angles are greater than 10 degrees, and by "bottomset slopes" and "bottomset beds" those parts below the foreset slopes where the dip angles are lower than 10 degrees. From the sedimentological point of view it would then be necessary to further subdivide the bottomset slopes and beds.

Sediments with median diameter of about 0.18 mm are the best sorted (*Inman*, 1949) and are often characteristic of beaches, river-mouth bars, and deltaic foreset slopes. Material with an equivalent grain size of 0.1—0.5 mm is more

easily eroded than any other equivalent grain size (*Hjulström*, 1935, 1939; *Sundborg*, 1956). The critical velocity at which sand movements start, owing to the frictional drag force of the fluid, is called the threshold velocity (*Bagnold*, 1941, p. 86). For spherical material with a density of 2.65 g cm^{-3} the settling velocity is higher than the threshold velocity, if the grain size is larger than 0.18 mm, and lower, if the grain size is smaller than 0.18 mm (*Inman*, 1949).

Particles larger than 0.18 mm may therefore have a somewhat different pattern of transportation and deposition than smaller particles. Particles smaller than 0.18 mm are usually transported as suspended load and particles larger than 0.18 mm both as bed load and as suspended load. The boundary between bed load (contact load), saltation load, and suspended load is rather diffuse and varies with the flow conditions (see *Sundborg*, 1963, Fig. 11, p. 65). The approximate boundary between bed load and suspended load may therefore differ from time to time and from point to point of a watercourse. The grain-size distribution of river-mouth bars and deltaic foreset slopes, however, indicates that here the diffuse boundary between bed load and suspended load may approach the grain size for which the settling velocity is equal to the threshold velocity. From the quantitative point of view the term "potential bed load" will therefore, for the sake of simplicity, mean particles larger than 0.18 mm, even if they are often transported as suspended load or as saltation load. However the term "bed load" without attribute here means contact load (rolling and sliding particles).

For the formation of foreset slopes bed-load transportation is necessary. Material transported as bed load usually comprises a small percentage of the total sediment transported. The normal percentage appears to be of the order of 10 % (cf. *Lane and Borland*, 1951). At the distributary mouths on large, gently inclined deltas, the percentage is often much lower. Foreset slopes are therefore usually of a fairly inconsiderable extent at the front of large deltas.

The extent of foreset slopes usually increases with the grain size of the transported and deposited material and especially with the ratio of bed load to suspended load. Their height is also dependent on the depth conditions in front of the distributaries, and they may be rather high at the places on the delta front where the original inclination of the basin floor is steep. The rate of deposition close to the foreset slopes and therefore also the rate of reduction in water depth decreases with decreasing grain size of the suspended load and the sediment may be rather widely dispersed when the inflow is of overflow type. This may therefore also influence the size of the foreset slopes (cf. *Axelsson*, 1956).

Waves and currents often have a destructive effect on deltaic foreset slopes. When bed-load transportation ceases, the height of the foreset slopes will be systematically reduced by deposition of suspended load in the transitional zone between foreset and bottomset slopes. This transitional sedimentary link (Fig. 3) is defined as a toeset deposit by *Jopling* (1960).

The dip angles of foreset slopes are usually somewhat lower than the static subaqueous angle of repose. The extensive literature on the subaerial angle of repose is summarized by *van Burkalow* (1945). The angle, usually of the order of 30—35 degrees, varies slightly with the grain size, the degree of sorting, sphericity and roundness, angularity, surface texture of the particles, density, and the surface curvature of the underlying slopes. For well-rounded, sand-sized materials, the angle of repose may be about 5 degrees lower than for subangular gravelly sand. According to the experiments by *Thoulet* (1887), *Piwowar* (1903), and *Jopling* (1960), the difference between the static subaqueous and the subaerial angles of repose is rather slight, but, according to *McKee* (1953, 1957), the subaqueous angle of repose is about 5 degrees lower than its subaerial counterpart. Foreset dip angles may also be influenced by post-depositional deformation, due, for instance, to seepage and consolidation processes.

The presence of clay and silt reduces the angle of the foreset slopes and increases the development of bottomsets (*McKee*, 1957). Sporadic mass movements, such as slip and slump of foreset material, with concomitant variations in dip angles, often characterize the foreset slopes. See the experimental investigations by *Jopling* (1960) and by *Johansson* (1963). In these laboratory studies the dip angle was somewhat higher before than after a major slump. Mass movements may also cause an increase in downslope grain size and a basal concentration of coarser grain sizes (see, for instance, *Shrock*, 1948, p. 246). This downslope variation is due to the dispersive pressure, generated during slip (see further *Bagnold*, 1941, 1954).

The dynamic angle is often lower but may temporarily exceed the static angle of repose. Increasing velocity of flow (bed tractive stress), depth ratio, and ratio of suspended load to bed load decrease the dip angles of foreset slopes and favour the transition from an angular to a tangential type of basal contact (cf. *Jopling*, 1960, 1963, 1965).

The height and the dip angles of foreset slopes may therefore be rather low, the contact with topset and bottomset slopes strongly tangential, and the grain-size contrast between topset, foreset, and bottomset slopes rather slight at the front of those distributaries that are built out in shallow basins, especially if the inflow is of turbidity current underflow type and/or the wave or tidal action is strong. The distinction between topset, foreset, and bottomset slopes may then be less significant, even if bed-load transportation is occurring.

Well-defined topset, foreset and bottomset slopes may be more common at the front of those distributaries that have a high ratio of bed load to suspended load and that are built out into deep basins, especially if the inflow is of overflow type and the wave and tidal action is weak.

It must, however, be emphasized that real foreset slopes, as here defined, are of a fairly inconsiderable extent in most of the large deltas of the world. *Shepard* (1956) uses the term "pro-delta slope" for the slope beyond the shallow-water platform of the Mississippi delta. For those deltas and those parts of a delta that have no real foreset slopes it would perhaps be better to use the term "delta-front slope" for the steeper part of the bottomset slope bordering the delta-front platform and the term "pro-delta slope" for the lower part of the bottomset slope.

The bottomset slopes, particularly the pro-delta slopes, are usually very stable in relation to the foreset slopes. Mass movements may, however, occur also on rather gently inclined slopes, especially if the rate of deposition is very high. On rapidly advancing delta-front slopes the shear strength of the sediments may be very low in comparison to other, more slowly deposited, sediments, as discussed by *Moore* (1961). According to *Terzaghi* (1956) the deposition of silt may sometimes occur in such a way as to cause the deposit to be in a metastable condition. Clay sediments may accumulate so fast that the processes of consolidation by reduction of pore water cannot keep pace, and deposits of this type are therefore called underconsolidated. The shallow gullies in front of the Mississippi delta, dissecting parts of the delta-front slopes with an average inclination of only 1 per cent, are cited by *Terzaghi* as probable examples of slope failure in underconsolidated sediments (cf. p. 23). The shear strength tests of delta-front sediments from the Mississippi delta, described by *McClelland* (1956) and by *Fisk and McClelland* (1959), confirm the underconsolidated condition.

Rapidly advancing delta-front slopes may thus be rather unstable, also if they are only gently inclined. Mass movements, such as subaquatic slumps, generated for example by earthquake shocks, may therefore sometimes influence the morphology of the delta-front slopes. The often simultaneously generated turbidity currents may at least affect the transportation and deposition of sediment both on the delta-front slopes and the pro-delta slopes, and they are therefore also of morphological significance.

Advance and retreat of delta shorelines

The relation between advance and retreat of delta shorelines is primarily dependent on the ratio of supply to removal of sediment (by distributaries, waves, tidal and non-tidal currents, winds, ice, vegetation and fauna) and on the rate and direction of base-level variations. See further *van Straaten* (1960).

Some deltaic areas show at the present time an inactive character, while some are in a particularly active stage of development. The amount of sediment transported by the delta-building rivers is of vital importance for the rate of advance and development of deltas. In large rivers the transportation of sediment may reach very imposing dimensions. The average annual sediment load of the Mississippi, for example, is of the order of 500 million tons (*Holle*, 1952). The average areal rate of advance of the Mississippi delta, however, is rather slow in relation to other deltas with a sediment supply of a corresponding order of magnitude. This is primarily due to the rather rapid subsidence within the deltaic area (caused by downwarping, isostatic lowering, compaction and consolidation of sediment) and partly also to the rather small grain size of the sediment supplied.

The topography of the basin also influences the rate of delta advance. Those deltas that are built

out into deep basins usually advance at a rather slow rate. In shallow basins, on the other hand, the rate of advance may be very high, especially if the base level is sinking. Such is the case, for example, in the northern part of the Caspian Sea, where parts of both the Terek delta and the Volga delta are advancing at a rate of more than 100 metres annually (*Samojlov*, 1952).

The rate of advance or retreat usually varies considerably between different parts of the delta front. The rate of advance may be high at the mouths of larger, active distributaries, but between the active distributaries the removal of sediment by beach drift and longshore drift may be greater than the supply. There are data on this point from the Rhône delta, where, for instance, the annual advance at the mouth of the Grand Rhône distributary is about 30 metres, while the annual retreat of the beach farther westward (south of the Faraman lighthouse) is about 15 metres (*van Straaten*, 1960, p. 417).

With increasing length of the distributaries the slope of the energy gradient will be decreased, the limit of backwater will retreat further upstream, and the exposure to wave and current action may be increased. This favours shifts in the position and activity of the distributaries and the formation of subdeltas. The direction of principal advance may therefore be shifted to places with a greater slope of the energy gradient or with less exposure to wave and current action, and formerly advancing parts may instead be retreating.

The sediment supply may thus shift to another part of the deltaic area, and the abandoned part may be modified by compaction coupled with wave and current action. According to *Scruton* (1960), a delta cycle therefore consists of constructional and destructional phases. The constructional phase will end when erosion by waves and currents begins to dominate. During the destructional phase compaction of sediment goes on, the sorting process may be rather active in shallow water, finer-grained material may be removed, and coarser-grained residual may be concentrated into thin veneers, beach ridges, and barrier islands. Large deltas are often built up in a step-by-step manner. "Local delta construction is followed by partial destruction, and later, by another constructional phase" (*Scruton*, 1960, p. 82).

The rate of advance or retreat may thus vary considerably from place to place and from year to year. The calculation of the average annual rate of advance (or retreat) along the delta front therefore requires access to reliable early map material. Such material is available as regards the mouths of the Don, Po and Danube. The average annual advance for the Don delta is 10 m, for the Po delta 20 m, and for the Kilia delta of the Danube 27 m (*Samojlov*, 1952).

Deltas with a sediment supply of the same order may have very different rates of areal advance. The correlation between sediment supply and volume of deltaic sediment is much better, but one also has to consider the rate of organic deposition and the rate of compaction and consolidation of sediment.

Chapter 3

THE DELTA PLAIN

General

The morphological delta pattern may be very differently developed in different topographical and climatological environments. Besides the flow conditions several factors, such as waves, winds, ice and vegetation, as well as the initial topography of the basin often affect the deltaic features.

Waves are an important factor in the redistribution of deltaic sediments. Important eolian transport may also take place in many deltas (*Tricart*, 1956, 1961; *Walker and Arnborg*, 1964). Beach accumulations as well as wind dunes may therefore be common along the delta shores and they are also found in the interior of several deltas. In cold climatic regions the formation of ice-jams is an important factor producing flooding and crevassing (*Henoch, 1960*; *Mackay*, 1963). The binding and filtering effect of vegetation increases the relief and assists levee building (*Russell*, 1936; *Zonneveld*, 1959; *Dahlskog*, 1964; *Fosberg*, 1966). In shallow basins the channel pattern may be greatly influenced by initial topographic irregularities (*Axelsson*, 1964).

The subaerial part of a delta, however, is primarily a deposit built up and modified by fluvial processes. Generally the deltaic pattern is in the main the result of fluvial activity, and the morphological development of the delta plain during the constructional phase is largely dependent on the flow conditions and on the grain-size composition of the transported and deposited material.

As a result of the back-water effect delta advance is usually followed by aggradation and by a progressive sorting and segregation of the transported and deposited bed material (see further *Harrison*, 1953; *Lane*, 1955; *Kuiper*, 1960; *Sundborg*, 1964). On the whole the grain-size and the slope angle therefore usually decrease slowly in a downstream direction but may vary considerably locally, especially in a lateral direction.

On small coarse-grained deltas the average inclination of the subaerial part may be rather steep, up to several per mille, as discussed by *Moravetz* (1960). The inclination of large, fine-grained deltas is much lower and often in the order of 0.05 per mille (*Samojlov*, 1952). The slope of the water surface in the deltaic channels is also dependent on, *inter alia*, the water discharge and on the water stages in front of the channel mouths and may therefore vary considerably during the year. As an example of this we may mention the Kemijärvi delta, where the average slope of the water surface may be higher than 0.06 per mille at rapidly rising discharges but as low as 0.004 per mille at falling discharges (*Axelsson*, 1964).

The flow conditions and the grain-size composition of the transported and deposited material may vary considerably between different deltas and between different parts of the same delta. The deltaic pattern may therefore also vary considerably between different deltas and between different parts of the same delta as well as with the lapse of time.

The subaerial part of a coarse-grained delta is often characterized by a rather rapidly shifting and oscillating river system. The branching river pattern is usually highly braided. Numerous bars of bed-load origin deflect the flow and cause channel splitting and rejoining both on the delta plain and at the delta front. The channels are shallow and the banks often gently inclined. Levees are usually poorly developed. The vegetation is often sparse.

Small coarse-grained, rather steeply inclined deltas as well as their proximal continuation often have the appearance of an alluvial fan. Valuable information about the morphological development and the sedimentary characteristics of especially the more proximal parts of these deltas may therefore be found in the literature on alluvial fans and highly braided rivers, as, for example, in the papers by *Hjulström* (1952, 1953, 1958), *Blissenbach* (1954), *Yatsu* (1954, 1959), *Arnborg* (1955), *Leopold and Wolman* (1957), *Moravetz* (1960), *Chien* (1961), *Krigström* (1962), *Doeglas* (1962), *Fahnestock* (1963), to mention some recent studies.

The subaerial part of a fine-grained delta is characterized by a better fixed and therefore

more slowly shifting and oscillating river system than that of a coarse-grained delta. Emerging bars of bed-load origin are less numerous. Channel splitting takes place mainly at the delta front. The channels are deeper, the banks often steeply inclined. Levees are usually well developed and inter-levee basins with swamps and lakes of different types may be numerous. The vegetation is often dense. The branching river pattern of a large, fine-grained delta therefore usually has a low degree of braiding.

Channel shape

The size and shape of a river channel is generally considered to be a function of the volume and character of the flow and the sediment load, of the character of the materials making up the bed and banks of the channel and of the character and density of the vegetation. When studying the equilibrium conditions for channels in various situations it is therefore necessary to pay attention to the laws for erosion, transportation and deposition of sediment as well as to collect and analyse quantitative data concerning channel characteristics. The contemporary knowledge of the laws governing the erosion, transportation and deposition of sediment is summarized in the detailed studies of fluvial processes by *Hjulström* (1935) and by *Sundborg* (1956). Available results from the method of quantitative analysis of channel characteristics presented by *Leopold and Maddock* (1953), and later widely applied, are summarized and further discussed by *Leopold, Wolman, Miller* (1964).

Streams which have evolved to a state of equilibrium have been called "graded". The most adequate definition of the term "graded" is probably that of *Mackin* (1948, p. 471). He defined a graded stream as "one in which, over a period of years, slope is delicately adjusted to provide, with available discharge and with prevailing channel characteristics, just the velocity required for the transportation of the load supplied from the drainage basin. The graded stream is a system in equilibrium; its diagnostic characteristic is that any change in any of the controlling factors will cause a displacement of the equilibrium in a direction that will tend to absorb the effect of the change."

The equilibrium profile of a channel is to a great extent dependent on the conditions for vertical and lateral erosion. As is clear from the well-known diagrams by *Hjulström* (1935) and by *Sundborg* (1956) showing the relation between particle size and critical erosion velocity for uniform material, there is generally a minimum of the erosion velocity within the interval 0.1—0.5 mm of particle size. The critical velocity for erosion thus normally increases towards both coarser, noncohesive and finer, cohesive sediments. The critical velocity for erosion is also dependent on, *inter alia*, the grain-size composition and the degree of consolidation of the sediments, and one also has to consider the binding effect of vegetation. In general, however, particles of sand-size are the most easily eroded. Channels with beds and banks of noncohesive sands thus have a low resistance to erosion and their shape and position may therefore be rapidly changed due to varying flow conditions.

The cohesive forces are independent of slope. Banks built up of cohesive material may therefore become very steep due to undercutting by currents and often also due to wave action. In channels with cohesive beds and with little transportation of bed load sometimes very steep furrows may be excavated by the corrasive action of sand and coarser particles along structural joints or less resistant strata (see further *Arnborg*, 1957; *Sundborg and Norrman*, 1963; *Norrman*, 1964).

For frictional material the critical erosion velocity decreases with increasing angle of slope (see *Sundborg*, 1956, Fig. 18, p. 202). This favours the conditions for lateral erosion in frictional material and the formation of a flat and shallow equilibrium profile. Banks built up of purely frictional material therefore often have a slope lower than the static angle of repose. When the rate of bed-load transportation is high the bed configuration may be rapidly changed, and a broad, flat channel may then provide the conditions for a temporary equilibrium, as further discussed by *Sundborg* (1956).

Deltaic channels with cohesive banks and with little or no transportation of bed load are therefore deeper, usually have steeper banks and better fixed position and shape than deltaic channels with banks of sand and coarser material and with a high rate of bed-load transportation.

Deltaic channels with cohesive banks often show remarkably straight courses. The tendency to meandering is usually most common in the proximal part of a delta and may be rather

pronounced in channels with active transportation of bed load.

In channel bends the flow velocity is normally higher at the outside than at the inside of the bends and is thus asymmetrically distributed. If the radius of curvature is small the flow may separate from the inside and the tendency to helicoidal flow be rather pronounced if the channels are narrow. This favours the deposition on the inside and the erosion on the opposite side of the channels, as further discussed in many papers, dealing with meandering rivers. See, for instance, *Sundborg* (1956), *Leopold and Wolman* (1960).

Deposition on the inside of a bend tends to diminish the cross section of the channel, to increase the average flow velocity and with it the erosive forces acting on the opposite side. Erosion on the outside of a bend on the other hand tends to enlarge the cross section of the channel, to decrease the average flow velocity and with it to favour the deposition on the opposite side. These opposite acting forces, that vary with the flow conditions, therefore favour a lateral displacement of channel bends and the formation of asymmetrical cross sections. The rate of lateral displacement may be rather high and the asymmetry of the cross section rather pronounced in those channel bends, where transportation of bed load is occurring, since longitudinal bars then may be rapidly built out on the inside of the bends.

In deltas sharper bends are often found at the head of channel forks and the fluvial processes here may be rather similar to those in a meander bend. Deltaic channels characterized by lateral displacement and asymmetrical cross sections are therefore found not only at normal channel bends but also at the head of channel forks. Especially in deltas with a high rate of bed-load transportation the accumulation of bars, that deflect the flow, may also cause lateral displacement of rather straight channel courses.

The equilibrium conditions for deltaic channels are often rapidly changed due to extension, splitting and closing of channels. These processes, of fundamental importance for the development of deltas, will be discussed in the later part of this chapter and also in Part II of this paper.

Natural levees

Deltaic channels with cohesive, vegetation-covered banks and with a rather coarse-grained suspended load are usually confined by well-developed natural levees. Levees may be formed due to the accumulation of sediment by flood-waters along the banks of channels. They may also originate from longitudinal bars, often formed in areas with lateral flow separation, as, for example, at the distal ends of channels, at the inside of channel bends and at the head of channel forks. Several, more or less parallel levee ridges are often formed on the accumulation side in connection with the lateral displacement of a channel. Long, curving, closely fitting levee ridges, built on the inside of channel bends, are also called scrolls (see *Melton*, 1936).

Levees are asymmetrical and often lens-shaped in cross section. The levee crest is usually situated close to the channel, because the quantity of sediment deposited is normally greatest near the channel and decreases in a lateral direction. This lateral decrease in deposition of sediment, usually followed by a lateral decrease in grain-size, is more pronounced with increasing grain-size of the suspended load and is greatly assisted by the filtering effect of vegetation. Levees built up along channels with a dense bank vegetation and with a coarse-grained suspended load may therefore have rather steep backslopes. Especially those levees that are built up on longitudinal bars of bed-load origin may at the beginning have very steep but short backslopes, since these bars often are characterized by a steep accumulation side facing the adjoining bank, as further discussed by *Sundborg* (1956, pp. 288—291).

The lateral displacement of a channel is accompanied by a gradual retreat of the levee-bank on the erosion side. Levee-banks often become very steep because of undercutting, especially at the outside of channel bends and at the downstream part of channel forks, where slides often occur at the active erosion scarps. The shape of levees is also modified by slower mass movements, and by the formation of crevasses, which according to *Russell* (1936, p. 78) occur especially at the outside of bends. The deposition of sediment on levees along abandoned channels may be very insignificant, and they may therefore decrease in height due to compaction and consolidation of sediment, as further discussed by *Russell* (1936).

Levees normally decrease in height towards the channel mouths, where they may continue beneath the mean water level. Along active channels they are usually sooner or later built

up to about high water level. The frequency, as well as the duration of water stages exceeding the bankfull stage, as defined by the crest of well-developed levees, will then be rather low.

Wolman and Leopold (1957) determined from annual flood series the frequency with which the bankfull discharge is attained at a number of rivers in the United States. They found that the discharge will equal or exceed bankfull 2 out of 3 years on the average at stations, where the flood plain is clearly defined and its elevation accurately known (see also *Leopold, Wolman, Miller*, 1964, pp. 319—322). *Nixon* (1959) concluded from his study of rivers in England and Wales that the bankfull discharge represents on the average the 0.6 % point on flow-duration curves, determined from mean daily flows. This means that the average duration of discharges exceeding bankfull is about 2 days per year for these rivers.

In deltas the younger and lower levees are flooded more frequently and during longer periods than the older and higher levees. The higher levee crests are not usually flooded every year. In the Laitaure delta, for example, they are topped by flood-water only about 2 out of 3 years on the average, even if they are overtopped twice in some years. The frequency of water stages exceeding bankfull stage in the proximal part of this delta is thus about the same as *Wolman and Leopold* (1957) found for flood plains in general. The duration of water stages higher than the higher levee crests is very short in the proximal part of this delta. The bankfull discharge here represents on the average the 0.1 % point on the flow-duration curve (Fig. 48), which means that the average duration is less than half a day per year.

Inter-levee basins

While natural levees are the dominant land form in many deltas, shallow inter-levee basins often occupy the greater part of the deltaic area. They are formed at the front of those deltas, where the splitting and extension of channels is followed by the development of new, more or less divergent and convergent levee systems. As these levees advance and grow higher, inter-levee areas will be embayed and later cut off from the frontal lake or sea basin.

Larger inter-levee areas are often subdivided by minor levee complexes into smaller basins, that are usually the sites of marshes, swamps and lakes. The delta lakes may be partly filled up by miniature deltas, formed at the mouth of inflowing channels. In those lakes that have only one channel connection with the active distributaries the connecting channel will function as an inflow during rising river stages and as an outflow during falling river stages. In those lakes that are entirely cut off from channel connection with the distributaries the water stage may sometimes differ considerably from that in the distributaries, where the water-stage variations are more rapid.

The sediment deposited within the inter-levee basins is often very fine-grained and usually has a high organic content as compared with the sediment deposited within the surrounding channels, and this is especially the case in those basins, that lack channel connection with the active distributaries. The sorting and segregation of the transported and deposited bed material within deltaic channels generally results in decreased grain-size downstream. This downstream decrease in grain-size is, however, usually very slow as compared with the decrease found laterally from active channels to adjacent inter-levee basins. Sorting is therefore chiefly a transverse affair in many deltas, as pointed out by *Russell* (1936, p. 77), and this is especially the case in those deltas where natural levees with adjacent basins flank the channels.

Bar building and channel splitting

The formation of distributaries and inter-levee basins at the distal part of a delta, schematically illustrated in Fig. 6, may be briefly described as follows.

At the mouths of the distributaries where the flanking levees are low and poorly developed, the river water spreads laterally and mixes more or less with the water of the adjoining basin. The marginal diffusion and deceleration of the expanding flow results in a deposition of bed load and suspended load that often creates a broad concentric distributary-mouth bar (A_1, A_2). As the bars grow in size they will deflect and split the flow and often cause the channels to bifurcate (B_1, B_2, B_3, B_4). As the new channels extend basinward they may bifurcate again and sometimes channels that converge upon each other may merge. If the splitting, extension and merging or re-joining of channels is followed by the development of new, more or less divergent and con-

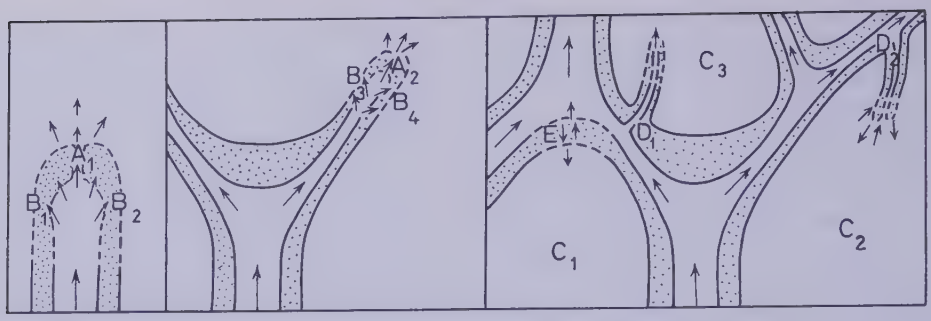


Fig. 6. Schematic formation of distributaries and inter-levee basins. (From *Axelsson*, 1955.)

vergent levee systems, inter-levee basins (C_1, C_2, C_3) may be embayed and later cut off from the frontal lake or sea basin. During high discharges minor overflow channels may be formed, widened and deepened at the outside of bends, and especially at the distal part of the channels where the levees are low and poorly developed and the vegetation is sparse. The majority of these channels are short-lived, some may evolve to larger distributaries while some may come to function as smaller channel connections (D_1, D_2) with the inter-levee basins. The passage at the distal part of embayed areas often remains open for many years after it has been cut off from direct connection with the frontal lake or sea basin. Sometimes this passage (E) develops into a channel that carries turbid river water into the inter-levee basin during rising river stages and clearer basin water out to the river channel during falling river stages. If the inter-levee basin has a deeper channel connection further upstream, the distal channel connection will mainly function as an outlet. According to *Mackay* (1963) the basin may then be subdivided by levees formed along the course of the through flow.

Bar building at distributary mouths, that often results from the marginal diffusion and deceleration of the expanding flow, may be very active when the rate of bed-load transportation is high (cf. *Mikhailov*, 1966). Waves are also frequently involved in the formation of distributary-mouth bars, as pointed out, for example, by *van Straaten* (1960, p. 413), and their shape may therefore be changed when wave action is vigorous. On marine deltas the bottom velocities may be low when salt wedges are present in the lower part of the distributaries and the transportation of bed load may therefore stop. According to the experiments by *Griesseier and Vollbrecht* (1953) the occurrence of salt wedges will also favour deposition of sediment and bar building because the vertical turbulent exchange will then be disturbed. Since fine-grained suspended particles will settle as flocculated masses in brackish and salt water, *Barton* (1918) assumed that the location of sandbanks and bars is chiefly controlled by the deposition of colloidal matter. *Bates* (1953, p. 2143) and *Welder* (1959, p. 54), on the other hand, are of the opinion that colloidal flocculation cannot be considered as important in bar formation, since most of the colloidal matter is removed by subsequent floods, and according to *Strickland* (1940) and *Santema* (1966) tidal currents will also help to keep the channels open. *Scruton* (1956) points out that the depth of the distributary-mouth bar is an equilibrium depth produced by a balance between forces tending to choke the channel and those tending to maintain it, and that bar depth bears a direct relation to distributary discharge. Distributary mouths therefore tend to choke up when part of their flow is diverted.

At the delta front the distributary channels are usually shallow and poorly defined. The channel mouths often lie close together and frontal bar deposits of active and abandoned channels therefore often merge laterally to form greater distal lobes. As the channel-mouth bar advances and the channel elongates the slope of the energy line tends to decrease upstream from the bar. This favours lateral migration or oscillation of the flow and later channel splitting at or upstream from the bar. The tendency to blocking and splitting of channels also increases with increasing exposure to stronger winds, waves and basin currents. Channel formation and

extension will therefore be successively shifted to places with a greater slope of the energy line or with less exposure to actual wave or current action. The initial route of a river channel may also be determined by the topography of the basin it crosses, as pointed out by *Russell* (1958, p. 12), and this is often the case in shallow basins where channel splitting may also be caused by initial topographic irregularities (*Axelsson*, 1964).

The conditions for erosion, transportation and deposition of sediment at the delta front are also influenced by several other factors and vary considerably between different environments and from time to time, as discussed in the preceding chapter. In general, however, blocking and splitting of channels, both at the delta front and further upstream, will take place more rapidly if the transportation of bed load is periodically strong than if there is only little or no transportation of bed load. This occurs because the bed load reacts more rapidly than the suspended load to local changes in the bed tractive stress and is more generally associated with the formation and migration of bars, that may increase in height during high water and emerge during low water.

Even upstream from the delta front, bar building and channel splitting may therefore be very active where the transportation of bed load is periodically strong. This tendency to channel splitting will then be especially marked if levees are missing or are poorly developed, banks are easily erodible, and if the discharge rapidly fluctuates. This is the case on many coarse-grained deltas. The conditions for bar building and channel splitting in supraaquatic environments are further discussed in several papers, dealing with highly braided rivers. See, for instance, the detailed studies by *Krigström* (1962) and *Fahnestock* (1963).

Channel closing

As the bars, which divide the flow, grow in height, they may be stabilized by vegetation. They will then be better able to fix the locations of points of current bifurcation. Later they often evolve into a horseshoe shape with the point of bifurcation more or less rounded and with levees formed along the channel margins.

With increased height the levees are better able to confine the flow and to resist channel splitting, and the equilibrium profile will then be deeper.

The resistance to flow caused by the bank at the point of bifurcation implies a slight damming action, that may be rather pronounced at high-water discharges, flattening the slope of the water surface upstream and steepening it in a lateral and downstream direction.

At the upstream part of a channel fork, where the banks usually are diverging, deposition often tends to dominate over erosion. Where the radius of curvature of the banks is small, the flow may separate from the inside, reverse eddies may be formed, especially at high-water discharges, and longitudinal, more or less curving bars may be built out.

Just upstream of the point of bifurcation, where the flow is mainly diverging, a central bar is often formed, having a slight proximal and often a steep distal slope, and a shape varying with the symmetry of the channel fork.

At and just downstream of the point of bifurcation, where part of the flow impinges against the bank and is deflected, the turbulence is heightened, and erosion often tends to dominate over deposition. At high-water discharges strong eddies are often formed and the tendency to helicoidal flow may be pronounced. The levee-bank will often be undercut, and deep, narrow pools may be excavated in the bed.

As previously discussed the shape of a river channel is influenced by a great many factors, and the shape of a channel fork may therefore vary considerably from place to place. An asymmetrical channel profile, however, often provides the conditions for a temporary equilibrium at the head of those branches that diverge at a great angle from the trunk stream, because the flow velocities here are usually asymmetrically distributed.

Natural channel forks may be symmetrical but they are more often asymmetrical with one of the branches smaller than the other and with one of the branches diverging at a greater angle from the trunk stream. Usually the less efficient branch is smaller and diverges at a greater angle. The size of the branches and the shape of the channel fork may also vary considerably with time, and often the less efficient branch will gradually be closed at the head.

Since the frictional resistance per unit volume of flow increases with decreasing size of a channel, other factors being equal, a channel loses efficiency

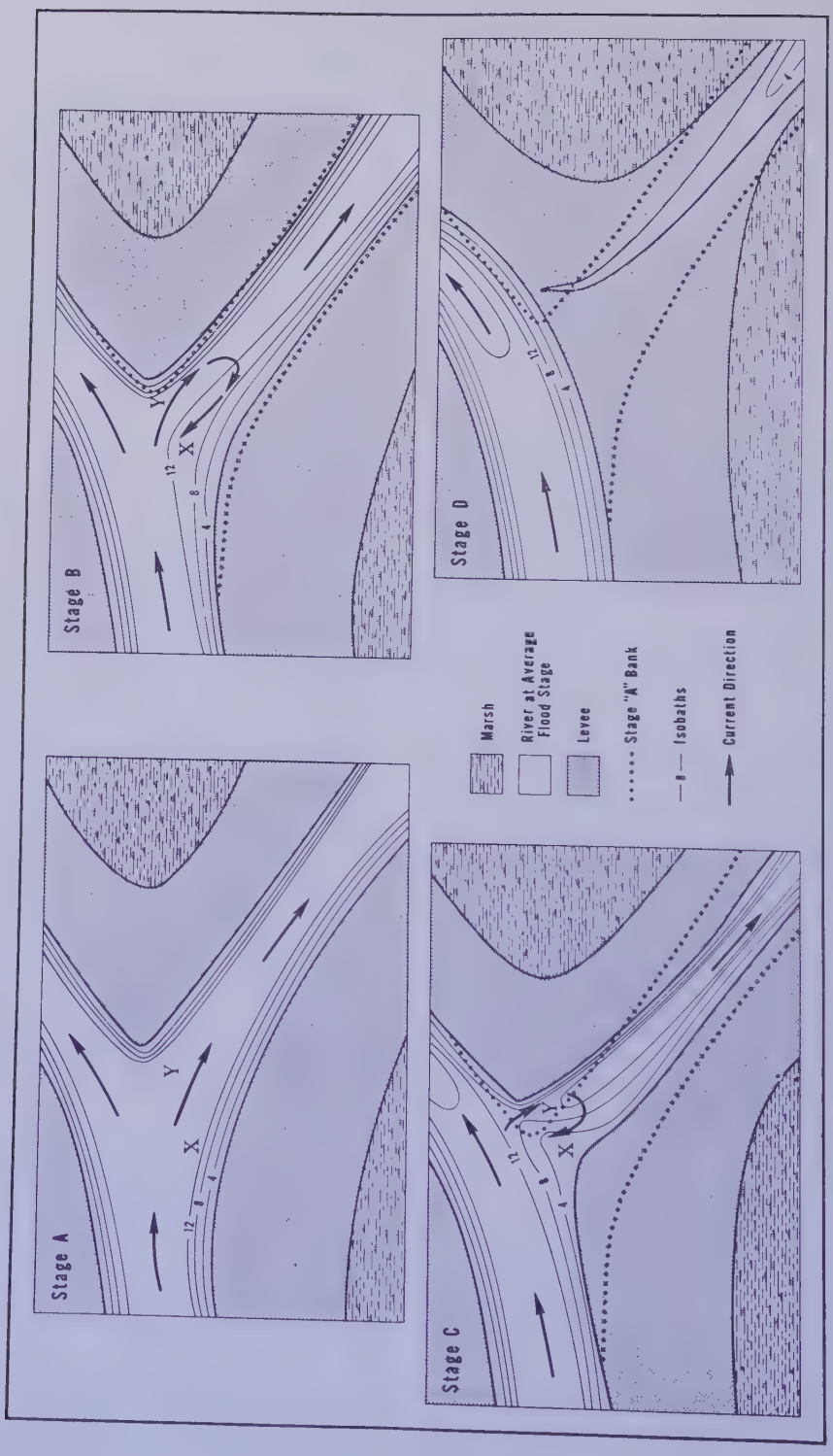


Fig. 7. Schematic development of channel closing. (From *Welder*, 1959.)

when it splits up into two branches (cf. *Welder*, 1959). This loss in efficiency to transport especially the coarser part of the sediment load may be reduced by a compensatory change in other variables. A new equilibrium profile may then be established, and according to *Leopold and Wolman* (1957) a divided reach is usually steeper, shallower, and wider than an undivided reach carrying the same flow (cf. also *Mackay*, 1963). Channel bifurcation may also be followed by channel closing and by a concentration of flow to one of the branches. The remaining active branch will then increase in size, as a result of the added discharge.

It is often assumed that a large angle of bifurcation from the trunk stream is one of the primary reasons for the ultimate abandonment of a channel (cf. *Russell*, 1936; *Fisk*, 1952; *Welder*, 1959; *Kolb*, 1963). This assumption is mainly based on the results of several experimental investigations, showing that under certain test conditions most of the bed load will be diverted into a channel branching at a high angle from the trunk stream, as a result of helicoidal flow phenomena (see, for example, *Bulle*, 1926; *Benini*, 1952). In general, however, this is not the case when the bed at the channel fork has attained an equilibrium shape, as will be further discussed in Part II.

As previously discussed the slope of the energy line and the transporting capacity of a channel tend to decrease as the channel elongates. Channel extension therefore favours aggradation, channel splitting, and the diversion of flow to new branches having a shorter course, and consequently a greater slope (cf. *Visentini*, 1953). The equilibrium conditions may also be changed by base-level variations. Even a slight damming, produced, for example, by the effect of wind, may lead to a blocking and closing of the branches which debouch in the actual or prevailing wind direction.

The development of channel closing, discussed in detail by *Welder* (1959), is schematically illustrated in Fig. 7, and may be briefly described as follows for channels confined by natural levees.

During the closing process the size of the entrance to the closing channel decreases by deposition of sediment, espccially at the upstream part of the entrance. As previously discussed, deposition of sediment on the inside of a channel bend favours erosion on the opposite side and therefore a lateral displacement of a channel. At the head of a closing channel, however, deposition of sediment at the upstream part of the entrance (at X in Fig. 7) usually takes place more rapidly than erosion at the downstream part (at Y) as the efficiency of the channel decreases. As filling proceeds, the angle of divergence between the closing channel and the trunk stream usually increases, as pointed out by *Welder* (1959), and gradually the closing channel may be completely dammed off by a natural levee, formed across its head. The lag between the requirements of the flow and the channel cross-section is great if the supply of available sediment is small (*Russell*, 1957, pp. 172—173), and this is especially the case downstream of the head of an abandoned channel, where deposition of sediment usually takes place very slowly. Abandoned channels therefore may remain open for a long time after they have been sealed off from normal flow.

The closing of one branch at a channel fork usually demands that the other branch is or may become capable, by erosion, of holding a larger flow. Since this is often the case, channel closing may take place when the equilibrium conditions are changing, and the competency and capacity of one branch decreases or increases in relation to the other.

Channel splitting will therefore often be followed by channel closing and by a concentration of flow within the older parts of a fine-grained delta to fewer, active distributaries. During the constructional phase these tend to become larger and larger at the expense of the smaller and less efficient ones.

Chapter 4

STRUCTURE

General

The usual text-book picture of delta structure as a series of topset, foreset, and bottomset beds, Fig. 8, comes originally from *Gilbert* (1885, Fig. 4), who described the structure as follows (pp. 106—107):

> "As the delta is built lakeward, the steeply inclined layers of the delta face are superimposed over the more level strata of the lake bottom and in turn come to support the gently inclined layers of the delta plain, so that any vertical section of a normal delta exhibits at the top a zone of coarse material, bedded with a gentle lakeward inclination, then a zone of similar coarse material, the laminations of which incline at a high angle, and at bottom a zone of fine material, the laminations of which are gently inclined and unite by curves with those of the middle zone.
>
> The characters of the fossil delta, or the delta as it exists after the desiccation of the lake concerned in its formation, are as follows: The upper surface is a terrace with the form of an alluvial fan. The lower slope or face is steep, ranging from 10° to 25°; it joins the upper slope by an angle and the plain below by a gentle curve . . . The structure as seen in section is tripartite. In the upper division the lines of deposition are parallel to the upper surface of the delta; in the middle division they are parallel to the steep outer face, and in the lower division they are gently inclined. The separation of the middle division from the lower is obscure. Its separation from the upper is definite and constitutes a horizontal plane."

Gilbert's classical description may be regarded as a rough but good generalization of the principal structural features of many coarse-grained deltas built out into deep water by rapidly shifting and oscillating braided-river systems. For grouping the deposits of these deltas the terms "topset, foreset and bottomset beds" are often very useful.

The principal structural features of fine-grained deltas differ considerably, however, from those of coarse-grained deltas. For fine-grained deltas and therefore also for the majority of large, modern deltas the terms "topset, foreset and bottomset beds" often lack utility, and ought to be replaced by other terms for the different environments of deposition, as previously discussed in Chapter 2. Cf. also the minor sedimentary structures, exemplified by photographed cores from the Mississippi delta in Fig. 9, with the usual text-book picture, Fig. 8.

Deltaic deposits have in recent years been the subject of increasingly intense study. Amongst the more detailed studied, larger, modern, marine deltas are those of the Mississippi (*Russel and Russel*, 1939; *Fisk*, 1944; *Fisk, McFarlan, Kolb and Wilbert*, 1954; *Fisk and McFarlan*, 1955; *Scruton*, 1955, 1960; *Shepard*, 1956, 1960; *Moore and Scruton*, 1957; *Shepard and Lankford*, 1959; *Lankford and Shepard*, 1960; *Coleman, Gagliano and Webb*, 1964; *Coleman and Gagliano*, 1964, 1965), the Rhône (*Kruit*, 1955; *Razavet*, 1956; *Razavet and Kruit*, 1957; *van Straaten*, 1959 b), the Fraser (*Johnston*, 1921, 1922; *Mathews and Shepard*, 1962), the Orinoco (*Van Andel and Postma*, 1954; *Nota*, 1958), and the Niger (*Allen*, 1964, 1965), to mention some recent sedimentological studies.

The works on these and on other deltas demonstrate that modern deltas show a wide variety in sedimentary characteristics and that our knowledge of their structural features is still rather incomplete. In the following a brief generalizing description is given of the more common deposits of various deltaic environments.

Delta plain deposits

Channel deposits

Channel deposits often comprise a wide range of grain-sizes which are more or less well sorted into a great variety of minor sedimentary structures. In active deltaic channels the grain-size of

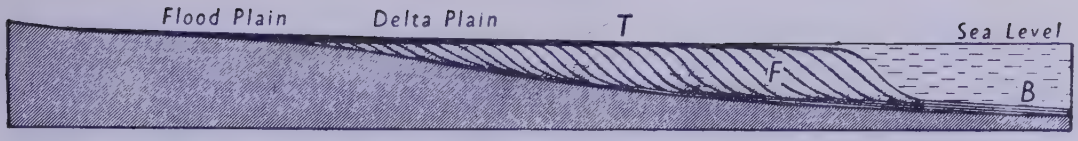


Fig. 8. Schematic section through the sediments of a Gilbert-type delta. *T*, topset beds; *F*, foreset beds; *B*, bottomset beds. (From *Holmes*, 1960.)

the bed material on the whole decreases slowly in a downstream direction but may vary considerably locally. Lateral and vertical relations are both gradational and sharp, and normally the coarsest material is found in the deeper parts of the active channels.

The transportation and deposition of bed load in the form of current ripples and bars gives rise to different types of cross-stratified units. Scour and fill structures are often formed due to rapidly varying flow conditions. Erosional truncation and enrichment of coarser residual material often

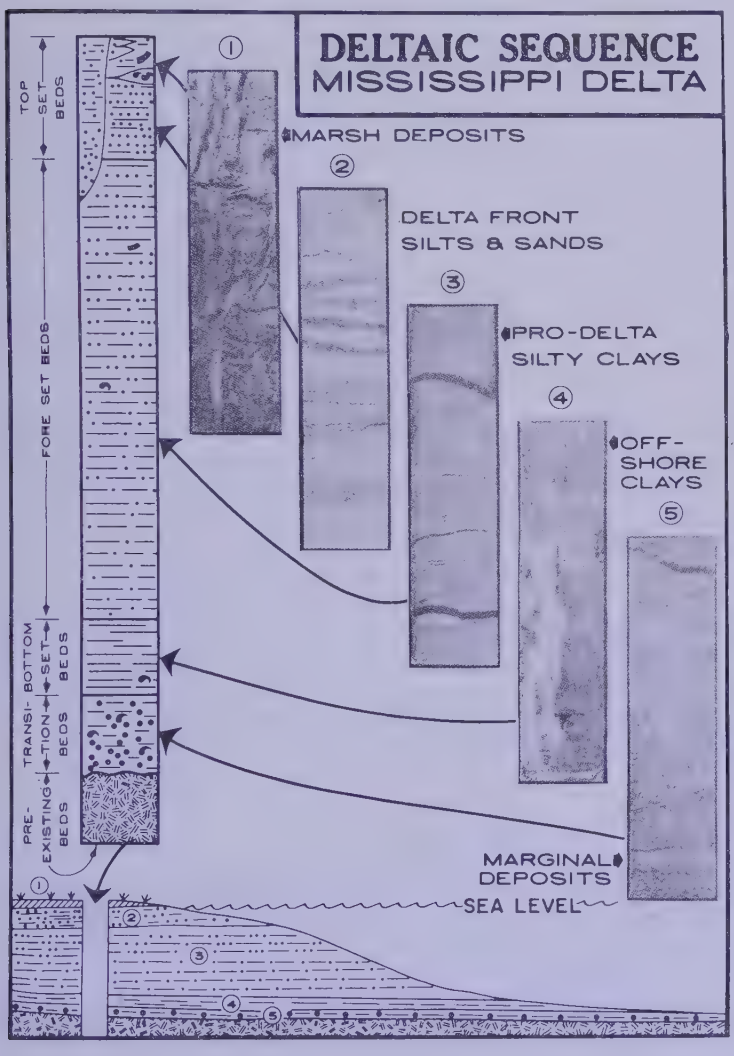


Fig. 9. The vertical deltaic sequence presently forming in the Mississippi delta. The vertical scale of the profile is greatly exaggerated. (From *Scruton*, 1960.)

take place during rising river stages, and sometimes upper parts of underlying delta front deposits may be eroded and replaced by coarser channel lag deposits.

Deposition in active channels generally occurs through the growth of point-bars and channel-head bars in connection with lateral channel displacement and channel closing. In sandy channels these bars often contain a well sorted, and mainly compound cross-stratified basal portion of bed-load material, capped by a deposit of mostly finer-grained, suspended material, which may be irregularly laminated as a result of deposition over ripple marks and scour structures. These bar deposits are often underlain by coarser lag deposits and they normally grade upwards into finer-grained levee deposits.

The rate of deposition may be rather high in channels undergoing gradual abandonment. In rapidly abandoned channels the aggradation is slower, the introduced fill is finer grained, and consists on finer-grained deltas mostly of laminated sandy and clayey silts. The upper, finer-grained parts of these channel-fill deposits may have a marshy character, and they often show a laminated structure, disturbed by plant roots and by burrowing animals.

Levee deposits

Natural levee deposits are on the whole finer grained than most channel deposits. They are normally characterized by alternating, often somewhat distorted, coarser and finer layers, generally containing drifted and autochthonous plant remains, and consisting mostly of fine sand and silt grading out to clayey silts down the back slopes.

Their often somewhat irregularly laminated structure may be partly disturbed by plant roots and by burrowing animals. As a result of subaerial exposure during the greater part of the year levee deposits often achieve a red-brown mottled appearance because of oxidation.

Inter-levee deposits

The grain-size of suspended sediment deposited from floodwaters decreases with increasing distance from the active distributaries. Within the marshes and the swamps the rate of deposition of inorganic material is usually very slow, and the deposits are normally very fine-grained, poorly sorted and have a high content of organic material as compared with most underlying platform deposits. They often consist of indistinctly layered or homogeneous, dark-colored clayey silts and silty clays, usually of a peaty character.

The crevasse-splay deposits and the deposits of those inter-levee lakes that have channel connection with the active distributaries are normally coarser and more distinctly layered.

Delta front deposits

Platform deposits

The term "platform deposits" here means those deposits within the nearshore area (above the break-point) including distributary-mouth bars, interdistributary bays and beaches.

These deposits are much more variable in their characteristics than underlying slope deposits. They often show a great variety of minor sedimentary structures, and they are subjected during a shorter or longer time to reworking and winnowing by waves and currents.

The finer-grained deposits are found on the more sheltered parts of interdistributary bays and lagoons and on the outer, deeper part of the platform. Distributary-mouth bars and beach accumulations are coarser, often sandy, and they often show a high degree of sorting.

The finer-grained deposits are mainly horizontally laminated, but the coarser bar deposits often compound cross-stratified. Wave ripple structures are widespread. Scour and fill structures are especially common in the distributary-mouth bars.

The tendency to erosional truncation, removal of finer-grained material and concentration of coarser-grained residual into thin veneers, beach ridges, and barrier islands increases with increasing exposure to wave and current action and may be very pronounced during the destructional phase.

Frontal slope deposits

The rate of clastic deposition generally decreases with increasing distance from the active distributary mouths. Normally the median grain-size of the deposited material also decreases in a distal direction. The proximal parts of the frontal slopes are therefore usually steeper, and

consist of coarser-grained deltaic deposits than the distal parts.

Steeply inclined slopes of a real foreset character with cross-stratified sediments, mainly consisting of deposited bed-load material, are common at the front of coarse-grained deltas but are of a fairly inconsiderable extent at the front of fine-grained deltas, as previously discussed in Chapter 2. The proximal slope deposits of fine-grained deltas consist mainly of deposited suspended material. The often silty deposits are generally laminated as a result of variations in the supply of sediment. Wave and current ripple structures and slump structures are often found on the upper parts of the slopes.

The distal deposits of marine deltas generally lack laminations. They consist mainly of flocculated clayey silts and silty clays. These deposits often achieve a mottled appearance as a result of organic activity during and after deposition, and they often have a high content of shell remains, especially where the rate of clastic deposition is very slow.

Stratigraphic sequence

In a single deltaic lobe the vertical distribution of sediments is roughly the same as their order of occurrence in a distal direction out into deep water. They are normally finest and most uniform in depositional fabric at the base, and become coarser grained and more complexly stratified upwards towards the top, where wave-agitated platform deposits are capped by a complicated series of channel, levee, and inter-levee deposits.

Large deltas, however, are usually built up in a step-by-step manner, as further discussed by especially *Egorov* (1955), *Scruton* (1960), and *Coleman and Gagliano* (1964). Due to shifts in position and activity of the distributaries and to base-level variations, local delta construction is followed by partial destruction, abandonment, and initiation of a new cycle. Large deltas are therefore normally a composite of numerous lobes, with the younger lobes formed successively on top of, in front of, or in marginal depressions of the older lobes.

Part II

EXPERIMENTAL STUDIES ON THE DIVERSION OF SEDIMENT AT BRANCHING CHANNELS

Chapter 5

PREVIOUS EXPERIMENTAL WORKS

General

The morphological development of a branching river pattern is greatly dependent on the special conditions for erosion, transportation and deposition of sediment at the head of channel forks, but our knowledge about these conditions in naturally branching channels is still rather incomplete. Technical problems linked with the diversion of sediment, especially bed load, at channel forks have, however, been the object of several experimental studies, and proposed methods for preventing sedimentation at intakes of irrigation and navigation channels and of water-power plants have long received much attention. These and other problems concerning the distribution of the sediment carried by a waterway which branches, either naturally or artificially, were discussed at the 18th *International Navigation Congress* (1953).

The first two experimental studies on the transportation of bed load in a forked flume were performed at about the same time by *Rehbock* at the Technical University of Karlsruhe and by *Thoma* at the Technical University of Munich in connection with a sedimentation-problem at a power-plant on the Middle Isar River. According to *Rehbock* (1926) and *Thoma* (1923) these experiments showed that a side channel, rectangular in cross-section, diverting 50 per cent of the flow from a straight channel of identical cross-section, diverts abnormally high amounts of the bed load. *Rehbock* explained this as a result of helicoidal flow phenomena at the channel fork, causing most of the lower water layers that transmit the bed load to deflect into the side channel.

The experiments at Karlsruhe were continued by *Bulle* (1926) to determine the amount of water and bed load diverted by channels branching from straight channels at various angles. *Bulle's* experimental flume was designed to include five angles of bifurcation (30, 60, 90, 120, and 150 degrees). The length of the straight channel was 6.43 m, and the side channel took off at 2.5 m from its upper end. The length of the side channel was equal to that of the straight channel below the point of bifurcation. Both channels were given a rectangular cross-section, a width of 20 cm, and a bottom slope of 0.003. Adjustable weirs were employed as tailgates at the downstream ends of the two channels. Bed material (probably medium to coarse sand) was introduced at the upper end of the straight channel, and after a given time interval the amount deposited downstream of the channel fork was collected and measured.

Some of *Bulle's* often quoted observations may be summarized as follows:

In the side channel opposite the point of bifurcation a reverse eddy was formed. This eddy acted to reduce the effective area of the section and to accelerate the constricted flow. At the point of bifurcation the water surface was superelevated. The transverse slope of the water surface was greatest at the entrance to the side channel. The size of the eddy and the transverse slope of the water surface decreased when the corner at the upstream part of the entrance to the side channel was rounded off, and when bed material was introduced into the flume. For tests of short duration (45 minutes), with sharp edges at the channel fork, and with equal water discharges in the two branches the percentage of bed load diverted by the side channel was always very high. The least deposition of bed load was noted in the side channel when the angle of

bifurcation was 120 degrees, but even in this case 87.5 per cent of the total was carried into it and deposited. As illustrated in Fig. 10 for a bifurcation angle of 30 degrees the percentage of bed load diverted by the side channel decreased when the weir at the end of the straight channel was lowered and the volume and velocity of flow was thus decreased in the side channel and increased in the other branch. When the corner at the entrance to the side channel was rounded off the percentage of bed load it diverted was decreased (cf. also *Tison*, 1953). The experiments also showed that the percentage of bed load diverted by the side channel was lower during the end of tests of longer duration when a bar was built up at the entrance to the side channel, and the bed morphology at the channel fork was thus in better equilibrium with the flow conditions.

According to *Benini* (1952) the experiments performed at the University of Padua by *Christani* also showed abnormally high amounts of bed load diverted into the side channel of a forked flume. In all these experiments the angle of bifurcation was 26.75 degrees. The straight channel was given a trapezoidal cross-section and a bottom width of 28 cm, while the side channel was given a rectangular cross-section. In test series No. IX the width of the side channel was 20.6 cm and the grain-size of the bed material 1—1.5 mm. The results of this test series are illustrated in Fig. 10.

Vogel (1934) reported results of preliminary experiments conducted at the *U. S. Waterways Experiment Station* (1932) along the same general lines as followed by *Bulle* (1926). Both channels were given a rectangular cross-section, and a width of 2 ft. The angle of bifurcation was 30 degrees. Two kinds of material were used: fine sand from the Red River, to show the effects of material carried largely along the bed, and loess, to show the effects of silty soils carried in suspension. In relation to the water discharge of the respective branches, a greater proportion of the fine sand moved into the side channel. However, when loess was the transported sediment, the division of the material between the two branches was nearly in proportion to the water distribution.

According to *Lindner* (1952) similar experiments performed at Iowa State College by *Dancy* (1947) also gave the result that suspended sediment, as differentiated from bed load, divides at a channel fork approximately in direct proportion to the division of water quantities. At natural channel forks the division of fine-grained suspended material is usually proportional to the division of flow (cf. *Turnbull and Kolb*, 1953).

Vogel (1934) also reported results of subsequent experiments conducted at the *U. S. Waterways Experiment Station* (1933). In these experiments the angle of bifurcation was 30 degrees. Both channels were given a semicircular cross-section. The main channel had a 2-ft radius and the smaller side channel a 1-ft radius. The bottoms of both channels were laid on the same horizontal plane. Traps were installed at the downstream end of the channels. After each run the model was drained, the sand held in the traps was measured, and cross-sections were taken of the sand deposits in the flume. During these experiments the flow in the side channel decreased gradually and slightly until stability was reached, but thereafter the ratio of flow in the two branches remained practically constant. It was also noted that during the first hours of these experiments both channels filled with sediment to a point of stability before any large quantities were caught in the traps. When equilibrium had been reached, after about 30 hours of running, the subsequent tendency was for nearly constant quantities to be carried through to the two traps. As pointed out by *Matthes* (1933) these experiments thus demonstrated the importance of continuing the duration of any experimental run until a stabilized condition is reached as regards both flow and sediment movement. When coarse Polk Creek sand was the transported sediment, the stabilized condition showed the side channel to discharge about 36 per cent of the total flow, and about 65 per cent of the total bed load. In these experiments the slopes in the side channel were steeper than those in the main channel. According to *Vogel* (1934) these experiments were therefore representative of the conditions found in a natural river at a reach where a small branch leaves the main river at an angle of 30 degrees to rejoin it below in such manner that the branch is shorter than the main stream.

Eicke (1958) also continued several of his experimental runs until a stabilized condition was reached as regards both flow and sediment movement. In his experiments the test conditions were on the whole more representative of natural channel conditions than in the earlier

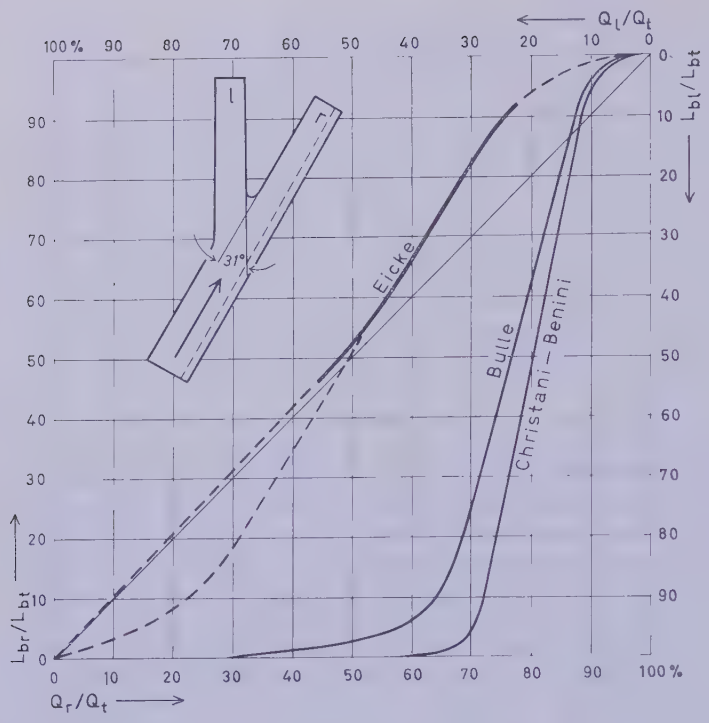


Fig. 10. Diversion of water and bed load at a channel fork according to experiments by *Bulle* (1926), *Christani-Benini* (1952) and *Eicke* (1958). Q_t = total water discharge; Q_l = water discharge in left channel; Q_r = water discharge in right channel; L_{bt} = total bed-load discharge; L_{bl} = bed-load discharge in left channel; and L_{br} = bed-load discharge in right channel. (Modified after *Eicke*.)

experiments referred to above. The experimental model was given a total length of 17—18 m and a bottom slope of 1: 600. The side channel was longer than the other branch. The corners at the entrance to the side channel were rounded. Traps were installed at the downstream end of the channels. The bottom of the model was capped by a sand layer, which was given a thickness of 22 cm along the channel sides and a thickness of 15 cm along the middle of the channels. Its initial thickness was thus enough to permit erosion and the formation of relatively deep pools at the point of bifurcation. The movable bed layer had a grain-size of 0.1—3 mm and a median grain-size of 0.7 mm. During the different runs sand of the same grain-size was fed in by hand at the upper end of the main channel. The water discharge was varied during the runs. At "high water" the maximum width of the water surface was 1.5 m in the main channel and 0.75 m in the two branches.

In *Eicke's* test series No. III the main channel was straight and the angle of bifurcation was 31 degrees. When the two branches had equal width the stabilized condition showed the side channel to discharge about 25—30 per cent of the total flow and about 15 per cent of the total bed load. When the width of the shorter branch was decreased to half the width of the side channel the stabilized condition showed the side channel to discharge about 45—50 per cent of the total flow and about 40—45 per cent of the total bed load. In Fig. 10 the results of this test series are illustrated and compared with those of *Bulle* and of *Christani-Benini*.

As is clear from Fig. 10 the results of *Eicke* differ considerably from those of *Bulle* and of *Christani-Benini*. Though boundary effects are exaggerated in small laboratory flumes the results of *Bulle* and of *Christani-Benini* indubitably demonstrate that curvilinear flow conditions, including flow separation and helicoidal flow, may to a great extent influence the diversion and the deposition of bed load at the head of a channel fork, when the morphology of the channel fork is not in equilibrium with the flow conditions. Curvilinear flow conditions thus influence the morphological development of a channel fork. *Eicke's* results indicate that the diversion of bed load at a channel fork is depen-

dent not on the angle of bifurcation but on the transporting capacity of the two branches, when the bed at the channel fork has attained an equilibrium shape. Observations on natural channel forks also indicate that the diversion of bed load as well as the diversion of suspended load is mainly dependent on the transporting capacity of the two branches (cf. *Hjulström*, 1952; *Arnborg*, 1955).

Exclusion of sediment from diversion intakes

In many artificial diversions the entrance to the branch channel is at a considerably higher elevation than the bottom of the main channel. Under these conditions little or no bed load will be diverted. Often, however, the sediment carrying capacity of artificial diversion channels is not enough to handle the sediment diverted into the head of the diversion channels (cf., for instance, *Inglis*, 1949; *Lindner*, 1952; *Lipscomb*, 1953; *Ismail*, 1954; *Riad*, 1961). Their capacity to divert water will then be reduced by deposition of sediment, especially at the head of the diversions.

The cost of keeping diversion channels in operation may therefore be very high. In Egypt, for instance, the equivalent of millions of dollars has been spent annually to remove sediment from irrigation channels diverting water from the Nile (*Leliavsky*, 1954). Proposed methods for reducing the amount of sediment entering the diversion channels have therefore received much attention.

To minimize the deposition of sediment at the entrance to diversion channels the ancient Arabs tended to place their intakes at the downstream end of the concave (outer) side of a bend, with the axis of the diversion tangent to that of the river. That the end of the concave side of a bend is a good location for a diversion intake was confirmed by several experimental investigations. See, for instance, *Habermaas* (1935), *Inglis* (1949), *Tison* (1953). The advantage of this location lies in the helicoidal flow, which tends to deflect the heavy fluid layer away from the concave side of a bend.

Artificial means, such as various types of sand traps, screens, sills, and vanes, are commonly used to reduce the amount of sediment, especially bed load, entering the diversion channels. As shown by model tests by *Martin and Carlson* (1953) favourable flow conditions can be obtained at the entrance to a diversion by the use of a curved guide wall. *Riad* (1961) found in his experiments that guide vanes directing the surface flow gave better results than dividing walls directing the bottom flow. *Tults* (1954) is of the opinion that the unbalanced centrifugal force, producing the helicoidal flow at the entrance to a diversion, may be balanced if the upstream embankment of an intake is given a greatly reduced radius towards the bottom. The helicoidal flow is utilized in *Nikitin's* curvilinear sand trap of continuous action. The water is led into a curved approach channel, smoothly bent through an angle of 90 degrees, with the bottom sloping from the concave to the convex bank. The transverse circulation arising in the bend carries the heavy fluid layer towards the convex bank, which is provided at the bottom with a number of openings, leading into discharge galleries. The material moving along the gallery openings is intercepted by them and is led downstream (*Nikitin*, 1954).

Several of the proposed methods for preventing sedimentation at intakes of diversion channels are summarized and further discussed by *Riad* (1961).

Helicoidal flow

The division of the flow at a channel fork varies with the size and the backwater characteristics of the two branch channels and with the morphology and the hydrodynamic conditions at the channel fork. The flow at the channel fork itself is complicated and may include helicoidal flow, flow separation, recirculation regions, hydraulic jumps, and both subcritical and supercritical flows (cf. *Taylor*, 1944; *Law and Reynolds*, 1966). Only subcritical flow conditions are considered in this paper.

As discussed in Chapter 3, flow separation and the formation of eddying regions is of importance for the morphological development of a channel fork. The more or less strong tendency to helicoidal flow that characterizes the flow pattern at channel forks as well as at channel bends in general is also of morphological importance. Helicoidal, or spiral, flow refers to movement of water particles along a helical path in the general direction of the flow. The transverse velocity components of the helicoidal flow will create so-

called secondary flow in the plane of the cross section.

Helicoidal flow also exists in straight channels (see, for instance, *Matthes*, 1947; *Einstein and Li*, 1958), but is more pronounced in curved channels. The existence of helicoidal flow at channel bends was described already in 1876 by *Thomson*, who explained it as being caused by the difference in "centrifugal force" between the faster flowing layers near the free surface and the fluid near the bottom, which has been retarded by boundary drag. This causes a helical motion with the top layers moving towards the outside of the bend and the bottom layers moving towards the inside. *Thomson* (1877) also constructed a small-scale model of a river-bend to demonstrate the phenomenon of helicoidal flow.

The centrifugal force is responsible also for the superelevation of the water surface along the outer bank of a bend. The total superelevation (Δy) over the whole channel width (b) is given by

$$\Delta y = \frac{1}{g} \int_{r_i}^{r_o} \frac{u^2}{r} dr \qquad (6)$$

where g is the acceleration due to gravity, u the flow velocity (constant, or varying with r), and r the radius of curvature. The subscripts, i and o, refer to the inner and outer radii of the bend, respectively. The shape of the transverse water-surface profile varies with the distribution of momentum. Several superelevation formulas have therefore been presented on the basis of various assumed conditions of flow (see further *Chow*, 1959; *Leopold et al.*, 1960; *Ippen et al.*, 1962). The overall bank-to-bank superelevation, however, is remarkably independent of variation in the radial distribution of velocities. Especially for channels of moderate curvature and of a large width-to-depth ratio (the norm in nature) the total superelevation is given with fair accuracy by the simple formula:

$$\Delta y = \frac{\bar{u}_s^2 b}{g r_c} \qquad (7)$$

where $\bar{u}_s$ is the mean flow velocity in cross section, and r_c the centerline radius.

Doubts as to the existence of helicoidal flow at natural channel bends have been expressed by, among others, *Vogel and Thompson* (1933, 1934), but in several field investigations the presence of helicoidal flow has been observed (see, for instance, *Blue, Herbert and Lancefield*, 1934; *Eakin*, 1935; *Rozovskii*, 1957). In natural channels, however, the transverse velocity components of the helicoidal flow are often of small magnitude. Especially in wide and shallow bends therefore the helicoidal flow pattern may be obscured by localized turbulence caused by channel irregularities (*Mockmore*, 1944).

The pattern of helicoidal flows in the bend of rectangular channels was studied by *Shukry* (1950), who used a specially designed pitot sphere to record the velocity components in the three planes. He found that the strength of the helicoidal motion (defined as the percentage ratio of the mean kinetic energy of the lateral currents to the total kinetic energy of flow at a given cross section) decreased with increasing Reynolds number of the flow (Re), with increasing ratio of depth to width (D/b), and with increasing ratio of centerline radius to width (r_c/b). The kinetic energy of the lateral currents was relatively small compared with the energy in the longitudinal currents. It attained practically its minimum value at $r_c/b=3$, and was nearly constant beyond this curvature ratio. According to *Bagnold* (1960) the total resistance to flow in a curved channel of uniform cross section also attains a minimum value at approximately the same curvature ratio (at $r_c/b=2$—3). When the channel bed was formed of erodible material *Shukry* found that the size of bed deformation decreased with an increase in r_c/b, and also that this bed deformation attained practically its minimum size at $r_c/b=3$. *Leopold and Wolman* (1960) found a large proportion of natural channel bends to have a curvature ratio between 2 and 3, with a median value of 2.7. These findings suggest that for subcritical flow, a curvature ratio between 2 and 3 should be recommended for a proper proportioning of curved artificial channels.

The results of theoretical, experimental and field investigations of the flow of water in bends of open channels, set forth by *Rozovskii* in 1957, are also of great importance for our knowledge of the flow conditions at channel forks. Assuming a logarithmic distribution of longitudinal velocities along the vertical *Rozovskii* arrived at the following formulas for the distribution of trans-

verse (radial) velocity components (u_z) in bends of wide streams:

$$u_z = \frac{\bar{u}\,D}{k^2 r}\left[F_1(y_r) - \frac{\sqrt{g}}{k\,C}F_2(y_r)\right]$$

(smooth boundary) (8)

and

$$u_z = \frac{\bar{u}\,D}{k^2 r}\left[F_1(y_r) - \frac{\sqrt{g}}{k\,C}F_4(y_r)\right]$$

(rough boundary) (9)

where $\bar{u}$ is the mean flow velocity in vertical, k the von Kármán "constant", C the Chezy coefficient, and y_r the relative distance above the bed, determined by the equation $y_r = y/D$.

The functions $F_1(y_r) = \int \frac{2 \ln y_r}{y_r - 1} dy_r$;

$F_2(y_r) = \int \frac{\ln^2 y_r}{y_r - 1} dy_r$;

and $F_4(y_r) = F_2(y_r) + 0.8(1 + \ln y_r)$ are graphically illustrated in Fig. 7 of *Rozovskii*'s paper. According to *Rozovskii* calculations by means of the formulas (8) and (9) show good agreement with data from laboratory and field investigations, except for the regions adjacent to the walls (banks).

Of great importance for the diversion of sediment at channel forks and for their morphological development as well as that of channel bends in general is the direction of the bottom currents, since this is the very flow direction that influences the direction in which channel-forming deposits are carried. For the central part of channel bends with a relatively small ratio of water depth to width and to radius of curvature (the norm in nature) one may use the following simplified formula for determination of the bottom flow deviation angle (φ) from the mean flow direction:

$$\operatorname{tg} \varphi = C\frac{D}{r} \qquad (10)$$

where the values of the constant C vary somewhat with the roughness, the distribution of longitudinal velocities, and with other factors. In general C has a value of 10—12 (*Rozovskii*, 1957).

In channel forks as well as in ordinary channel bends where the flow velocities are normally asymmetrically distributed, the shape of the channel cross-sections is usually asymmetric. The highest flow velocities and the greatest depths normally occur near the outside bank, and the bed is mainly sloping from the inside towards the outside. The deviation angle of the bottom currents from the mean flow direction varies with the ratio of water depth to radius of curvature. It will normally also vary both along and across the channels.

As regards the transport direction of bed load, one also has to consider that particles on a sloping bed will have a gravitational force component directed down the slope, equal to $m'g$ sine S, where m' is the immersed weight of the particles, and S the slope angle of the bed (cf. *Sundborg*, 1956). When this force component has attained a value equal to the transverse fluid-force component directed up the slope, no further net deflection of the bed load takes place towards the inside, which may probably be the case when the bed at channel forks and channel bends has attained a state of temporary equilibrium.

Chapter 6

EXPERIMENTS AT THE GEOMORPHOLOGICAL LABORATORY IN UPPSALA

Introduction

As pointed out in the last part of Chapter 3, it is often assumed that a large angle of bifurcation from the trunk stream is one of the primary reasons for the ultimate abandonment of a distributary. This assumption is mainly based on the results of some of the previously discussed experimental investigations, showing that under certain test conditions most of the bed load will be diverted into a channel branching at a high angle from the trunk stream, as a result of helicoidal flow phenomena. The author's observations on naturally branching channels, however, indicate that the diversion of bed load at a channel fork, as well as the closing of a distributary, is more dependent on the backwater characteristics of the respective branch channels than on their temporary angle of divergence from the trunk stream. It was therefore decided that the correctness of these observations should be tested experimentally.

Natural channel forks are usually asymmetrical. The highest flow velocities and the greatest depths normally occur near the outside bank, and one of the branches often diverges at a greater angle from the trunk stream. Changing equilibrium conditions are usually followed by aggradation or degradation. In order that the test conditions should be fairly representative of natural channel conditions it was therefore decided that the experiments should be carried out in a forked flume, whose bottom should be capped by a sand layer thick enough to permit degradation and the formation of asymmetrical channel cross-sections not only through accumulation along the inside but also through erosion along the outside of the entrance to its branch channels, and that one of its branch channels should diverge at a greater angle than the other from the approach channel.

Experimental equipment

The test flume

The experiments were conducted in the forked flume shown in Fig. 11 and 12. The flume consisted of a straight approach channel 1.0 m wide, and of two branch channels 0.5 m wide. The two branch channels were of equal length and diverged 50° to the right and 10° to the left from the approach channel. The flume had a total length of 8.4 m (measured along the centerline of the respective channels), and its bifurcation point was 4.0 m from the upstream end. The centerline slope of the respective channels was set at 1: 500.

Because part of the flume was to be used for subsequent experiments concerning the movement of water and sediment in curved channels, it could not be built as large as was desirable. For the same reason the distal part of the branch channels was given a centerline radius of curvature of 2.25 m.

A general description of the geomorphological laboratory is given by *Hjulström and Sundborg* (1962).

Water supply system

The water supply was recirculated from a large low-level tank from which it was pumped to a high-level tank equipped with horizontal, sharp-edged weirs that permitted the desired discharge to be drawn at constant head. The excess water not needed in the experiment flowed over the weirs and was returned to the low-level tank. The water drawn into the flume system was discharged through pipes from the high-level tank to the measuring tank shown in the background of the photograph, Fig. 12. The water discharge from this tank could be measured with an accuracy of about 1 per cent. The water was then led to a stilling basin at the upper end of the flume, and through a stone-filter to smooth out the turbulence before it reached the flume proper.

At the downstream end of the branch channels the water was discharged under and through perforated sluice-gates. Two sets of differently perforated plates which could be moved in relation to each other were used to vary the size and the number of effective sluice-gate perforations. The water stage and the flow

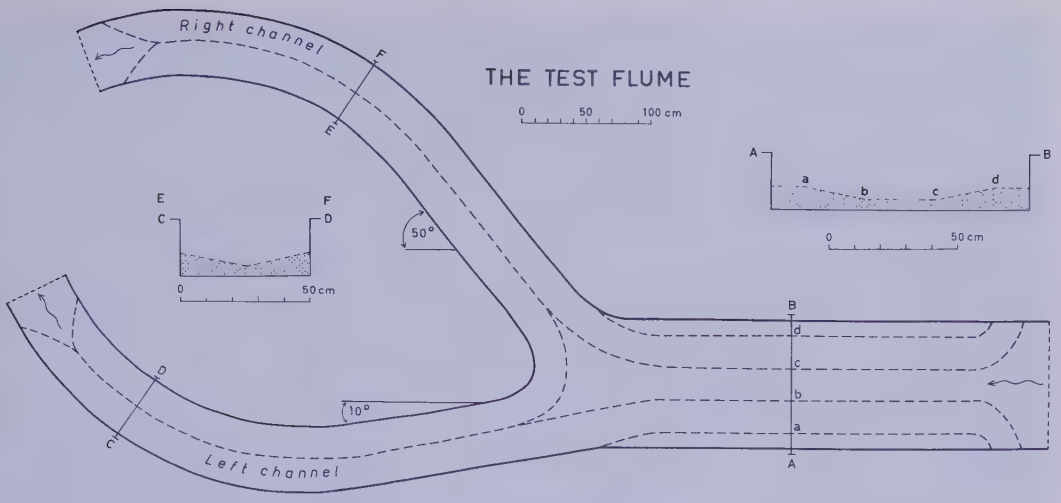


Fig. 11. Plan of the test flume. Initial thickness of the bed layer for run 1.

velocity could be varied by varying the water discharge, the distance between the sluice-gates and the flume bottom, and the size and number of effective sluice-gate perforations.

The basins at the end of each branch channel (shown in Fig. 12) were furnished with boxes for catching the bed load passing under the sluice-gates, with stone-filters to smooth out the turbulence, and with water-discharge measuring devices (point gauges and so-called Thomson weirs). From these measuring basins the water discharged through a floor well to the low-level tank.

The bed material

The flume bottom was capped by a layer of well sorted sand (a mixture of the Baskarp sands 3 and 5) having a median grain-size of 0.61 mm in diameter and a phi deviation measure of 0.67. The grain-size distribution of the bed material is illustrated in Fig. 13.

During each run dry sand of the same grain-size distribution was fed into the upper end of the approach channel at the desired rates by means of a vibrating feeder shown in the background of the photograph, Fig. 12.

The wet-volume of the bed material discharged at the end of the respective branch channels was measured upon the completion of each run. Later the discharged bed material was dried and its dry-volume measured.

Channel cross-sections were drawn by means of a point gauge that could be moved along the carriage shown in Fig. 12.

Experiments and results

Test series I

The aim of this test series was to study the discharge of bed load from the branch channels during about equal backwater conditions at the end of the left as at the end of the right channel. The sluice-gates were therefore set at the same elevation and given the same number and size of effective perforations.

Runs 1—13

After a few preliminary runs for testing the experimental equipment the sand layer was given a thickness of 9 cm along the channel walls, a thickness of 4 cm along the middle of the channels, and a mean thickness of 6.5 cm except for the entrance and the exit sections as shown in Fig. 11.

During each of the following thirteen runs (runs 1—13) the total water discharge was varied by steps as shown in Fig. 14. The rate of bed-load feeding (L_{bf}) was also varied by steps, and was highest (5 litres/30 minutes) at "high-water discharge" (30 litres/second). For each of these runs the entire bed-load feed was 10 litres.

Fig. 12. The test flume with connecting basins and some measuring devices. The photograph was taken after the completion of run 13

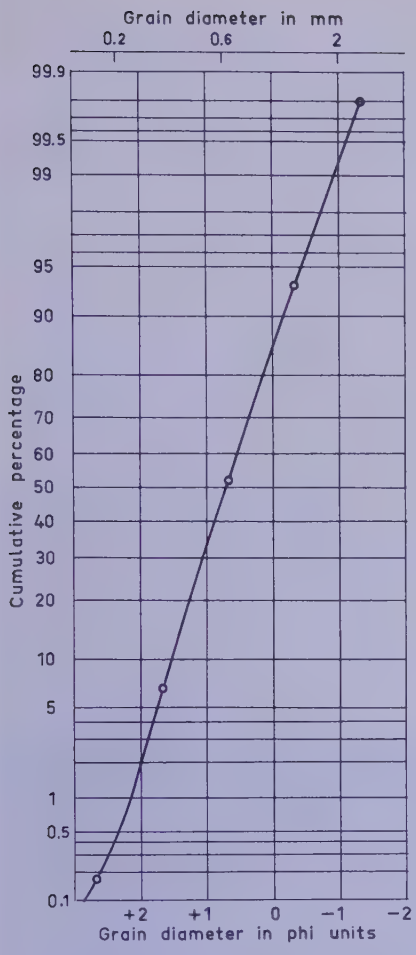


Fig. 13. Grain-size distribution of the bed material.

The duration of each of these runs was 3 hours. The water discharge from each of the branch channels was measured 10 times/run, and the water-surface elevation at selected sections 5 times/run (see Fig. 14).

When the total water discharge was held at 5, 15, and 30 litres/second respectively the average water depth was 2.0, 5.1 respectively 9.3 cm, and the average flow velocity 25, 29 respectively 32 cm/second. The average water discharge was about 6 (1—11) per cent higher in the right than in the left channel.

For run 1 the mean total bed-load discharge at the end of the branch channels was considerably higher than the mean bed-load feed. It was higher also for each of the following six runs though it mainly showed a decreasing tendency towards a value equal to the mean bed-load feed. After run 7 (21 hours of running) the difference between the mean total bed-load discharge and the mean bed-load feed was rather insignificant. For each of the runs 1—6 the mean bed-load discharge was higher from the right than from the left channel, but during the following seven runs about equal volumes of bed load were discharged from the left as from the right channel (see Fig. 15).

During the first seven runs, when erosion dominated over accumulation, the average thickness of the bed layer decreased with a decreasing rate from 6.5 cm to about 5.7 cm. During these runs a curved, elongated pool was excavated down to the flume bottom along the downstream wall of the channel fork (see Fig. 16), where the flow was superelevated, where especially at "high-water discharges" strong eddies and secondary currents were formed, and where deflection of bed particles towards the inside was observed to take place along the ripple-troughs. Accumulation took place along the inside of the entrance to each branch channel, where the flow tended to separate from the wall, and where reverse eddies were formed periodically. Asymmetrical channel profiles with pools along the outside walls and with less marked "point-bars" along the inside walls were also formed along the curved downstream parts of the branch channels.

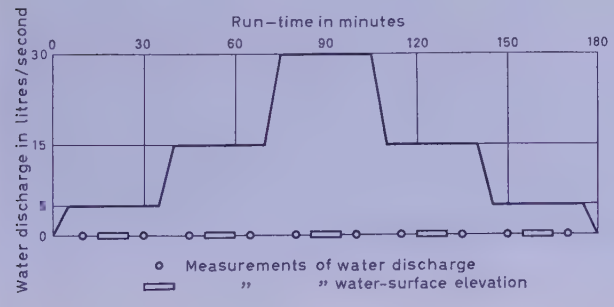


Fig. 14. Variation of total water discharge during each of the runs 1—13.

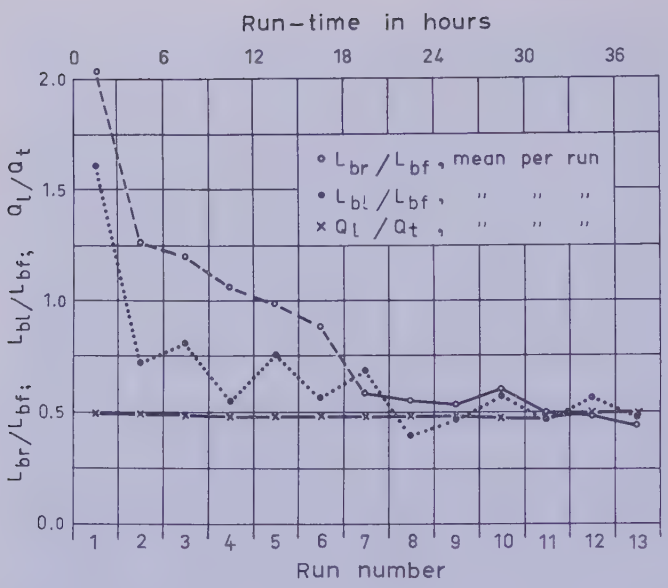


Fig. 15. Relative discharge of water and bed load from the end of the branch channels. Mean values for each of the runs 1—13.

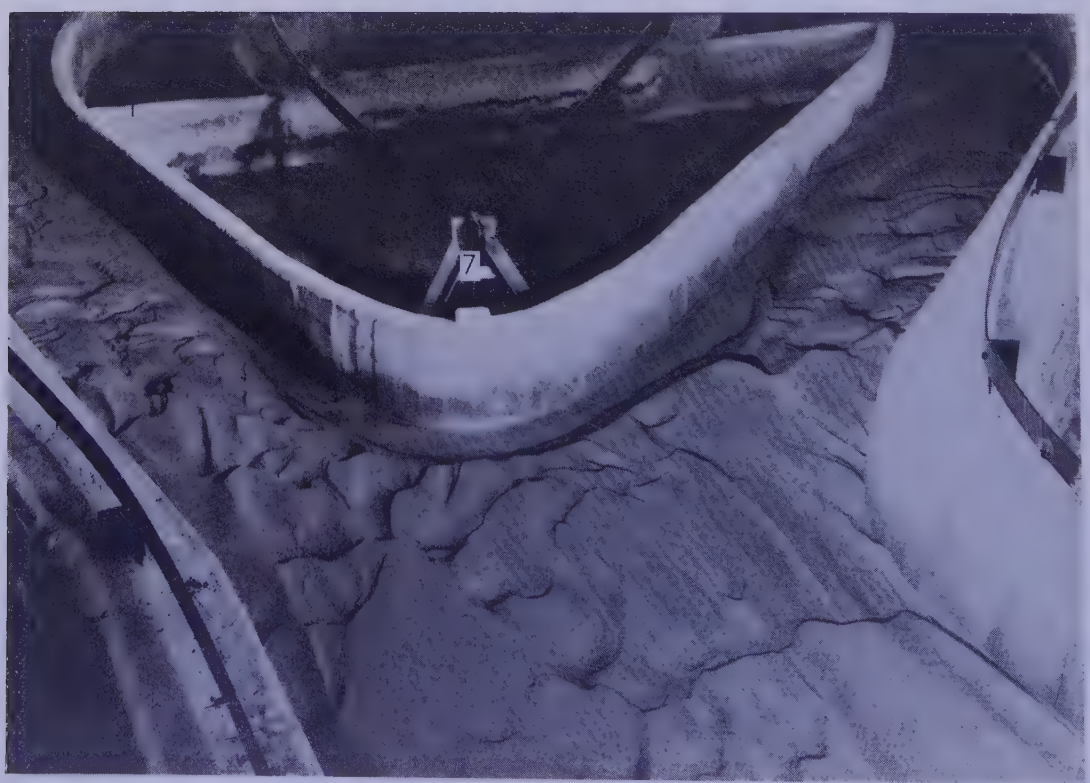


Fig. 16. The channel fork after the completion of run 7.

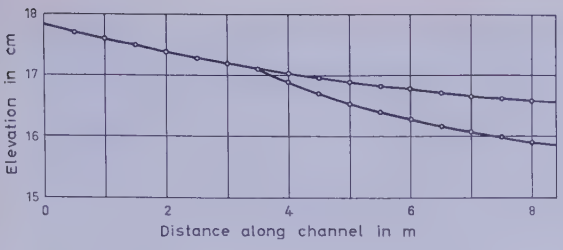


Fig. 17. Water-surface elevations along the channels. Mean values of the runs 20—24.

The size of bed deformation was largest at the entrance to the right channel, where also the flow velocities were most asymmetrically distributed. The crest of the "point-bar" formed at the inside of the entrance to this channel attained a maximum height of 12.7 cm above the flume bottom.

When equilibrium was established (after about 20 hours of running) the bed showed about the same configuration at the end of each run. The bed configuration, however, continued to vary somewhat with the water discharge during the runs. The pools attained their largest size during "high-water discharges". During "low-water discharges" parts of the "point-bars" were emerged.

Test series II

The aim of this test series was to study the discharge of bed load after the equilibrium conditions had been rapidly changed because of altered backwater conditions.

In four preliminary runs of different duration the total water discharge and the bed-load feed was held constant at 30 litres/second and 5 litres/30 minutes respectively (except for a short period at the beginning and at the end of each run). When equilibrium was established the average rate of bed-load movement was about equal in the left as in the right channel.

Runs 20—24

After the completion of the preliminary runs the distance between the sluice-gates and the flume bottom was increased by 8 mm (from 10 mm to 18 mm). The number of effective perforations was reduced by 33 per cent in the left sluice-gate, but was not altered in the right sluice-gate.

Also during the following five runs (runs 20—24) the total water discharge and the bed-load feed was held constant at 30 litres/second respectively at 5 litres/30 minutes. The duration of each of these runs was 3 hours.

For these runs the average slope of the water surface was steeper in the right than in the left channel (see Fig. 17). The average water depth was 9.1 cm in the approach channel, 8.8 cm in the right channel, and 9.4 cm in the left channel. The average flow velocity was 33 cm/second in the approach channel, 37 cm/second in the right channel, and 29 cm/second in the left channel. The average water discharge was about 14 (13—15) per cent higher in the right than in the left channel.

For each of these runs the bed-load discharge was considerably higher from the right channel than from the left channel (see Fig. 18). During the runs aggradation took place in the left channel. The bed-load discharge from this channel increased as the bed tended to be built up to

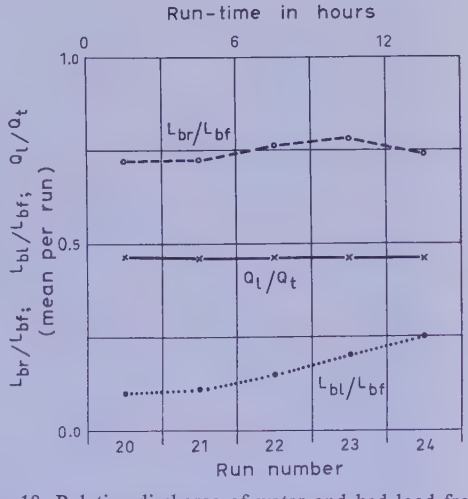


Fig. 18. Relative discharge of water and bed load from the end of the branch channels. Mean values for each of the runs 20—24.

Fig. 19. The channel fork after the completion of run 35.

Fig. 20. The channel fork after the completion of run 40.

grade. Degradation took place in the right channel. The degrading was, however, partly limited because the thickness of the bed layer was to small.

Test series III

The aim of this test series was to observe changes of the bed configuration occurring when lateral erosion took place at the downstream part of the channel fork.

After the completion of four preliminary runs, during which the test conditions were about equal to those during the runs 1—13, and during which equilibrium was established, the wall along the downstream part of the channel fork was replaced by a "cohesive bank" (a mixture of 77 per cent fine sand and 23 per cent clay).

Also during the following two runs (runs 34—35) the other test conditions were about equal to those during the runs 1—13, except that the duration was longer (6 hours/run). It was observed during these runs that lateral erosion took place along the "cohesive bank" when the total water discharge was held at 30 litres/second, that the rate of lateral erosion decreased during the runs, that the widening of the channel fork was accompanied by aggradation, and that the "point-bars" at the channel fork tended to increase in size. Also during these runs the bed-load discharge was about equal from the left as from the right channel. The channel fork after the completion of these runs (12 hours of running) is shown in Fig. 19.

After the completion of run 35 the distance between the right sluice-gate and the flume bottom was increased by 15 mm, while that between the left sluice-gate and the flume bottom was increased by 8 mm. The damming was therefore higher at the end of the left channel than at the end of the right channel during the following five runs (runs 36—40). Also during these runs (30 hours of running) lateral erosion took place with a decreasing rate along the "cohesive bank". Degrading occurred at the entrance to the right channel, while a tendency to aggrading was observed at the entrance to the left channel. The channel fork after the completion of these runs is shown in Fig. 20.

In some subsequent runs it was observed that the diversion of bed load at the channel fork could be made to vary considerably by changing the flow directions and the lateral distribution of flow velocities in the approach channel.

Conclusions

When applying these experimental results to natural channel conditions one has to consider, among other things, the limited size of the experimental model, the small thickness of the bed layer, and the lack of suspended load during the runs. Due to the limited size of the experimental model the boundary effects were exaggerated; because of the small thickness of the bed layer channel closing could not take place, since the closing of one branch usually demands that the other branch is or may become capable, by erosion, of holding a larger flow; due to the lack of suspended load levees could not be formed; etc.

The morphological changes that took place at the channel fork under different test conditions were, however, rather similar to those observed in naturally branching channels, which were further discussed in the last part of Chapter 3. "Point-bars" were formed at the inside of the entrance to the branch channels, "pools" were excavated at and downstream of the point of bifurcation, the size of bed deformation was largest at the entrance to the branch channel that diverged at the greatest angle from the approach channel. When the equilibrium conditions were changed and the competency and capacity of one branch was decreased in relation to the other, aggradation took place at the entrance to the less efficient branch, while degradation took place in the more efficient. Lateral erosion at the downstream part of the channel fork was accompanied by accumulation at the upstream part of the channel fork.

The experimental results showing that the diversion of bed load at a channel fork is more dependent on the backwater characteristics of the respective branch channels and on the flow conditions in the approach channel than on the temporary angle of divergence of the respective branch channels from the approach channel, may therefore be regarded as reliable.

The previously mentioned experimental results of *Bulle* and of *Christani-Benini* are often wrongly interpreted and wrongly applied. Their experimental results showed, *inter alia*, that accumulation will take place at the upstream part of the entrance to a rectangular side channel, because a rectangular channel shape here does not provides the conditions for a temporary equilibrium. Their experimental results, however, did not show

that most of the bed load will be diverted into a channel branching at a high angle from the trunk stream when the bed at the channel fork has attained an equilibrium shape, because equilibrium conditions were not established in their experiments.

One also has to consider that at natural channel forks the angle of divergence of a branch channel from the trunk stream may vary considerably with time, and that it often tends to increase when the efficiency of the branch channel decreases in relation to the other branch channel, due, for instance, to altered backwater conditions. At natural channel forks the less efficient branch therefore usually diverges at a greater angle from the trunk stream than the more efficient branch.

The morphological changes occurring at a channel fork reflect changes taking place upstream and downstream of the channel fork. The temporary angle of divergence of a branch channel from the trunk stream may therefore be regarded as a secondary factor.

Part III

THE LAITAURE DELTA

Chapter 7

GENERAL

Drainage basin

The Laitaure delta is situated at the eastern border of the Scandinavian mountain range in northern Sweden, just north of the Arctic Circle. It has been formed in Lake Laitaure by the river Rapaälven, which drains the central part of the high mountain region of Sarek. This river is a tributary to the river Luleälv through the Blackälven and Lilla Luleälv rivers.

Rapaälven has a total length of 74 km, and a length of 53 km upstream of the Laitaure delta (*Wersén*, 1925). The two largest tributaries are the Sarvesjokk and Kåtokjokk rivers. The downstream part of Rapaälven (downstream of Laitaure) is also called Laitätno. The map of the drainage basin (Fig. 21) is mainly based on the map of "Ortnamnen i Sarekfjälltrakten" (compiled by *Hamberg* in 1922) as regards the drainage net and the approximate height contours; on the map of "De svenska vattendragens arealförhållanden, Luleälv m. fl." (compiled by *Wersén* in 1925) as regards the basin boundary; and on the "Glacier map of Sweden" (compiled by *Schytt* in 1960 and revised by *Vilborg* in 1962) as regards the glaciers.

According to *Wersén* (1925) the drainage area comprises 814 km^2 in total, 670 km^2 at the Litnok hydro-meteorograph 3 km upstream of the Laitaure delta), 684 km^2 at the proximal part of the Laitaure delta, and 796 km^2 at the outlet from Laitaure. Lakes constitute 2.0 per cent of the total drainage area, but only 0.3 per cent of the drainage area upstream of the Laitaure delta.

The drainage basin is roadless and uninhabited except for the small habitation of Aktse. Most of our knowledge about this drainage basin comes from the scientific investigations, conducted by *Axel Hamberg*, who spent practically every summer during the years 1895—1931 in Sarek. The larger part of the drainage basin, however, is highly inaccessible, and is still rather poorly known both geographically and geologically. The available topographic maps (dating from the 1880's, on a scale of 1:200,000) are highly inaccurate. The above given area figures must therefore be regarded as approximate.

The drainage basin upstream of the Laitaure delta is part of the Sarek national park, which comprises the most glacierized high mountain region of Sweden. About 40 glaciers, having at present a total area of about 80 km^2 (about 1/4 of the total glacier area in Sweden), are drained by tributaries to Rapaälven. Glaciers thus constitute about 10 per cent of the total drainage area, and about 12 per cent of the drainage area upstream of the Laitaure delta. The largest of these glaciers, the Pårte glacier, situated at Pårtetjåkko, has an area of 14.1 km^2, and is the largest glacier in Sweden (*Schytt*, 1960). The best known of these glaciers, the Mikka glacier, situated at Sarektjåkko, has an area of 7.4 km^2 (*Stenborg*, 1965).

The Sarek mountains are deeply dissected by numerous valleys. The smaller valleys often debouch hanging into the larger, often trough-shaped valleys. In the large, open valleys the watershed is at an elevation of about 900 m (*Hamberg*, 1901; 1910 b). Within and at the boundary of the drainage basin more than 30 peaks rise above 1,800 m. The highest of these, the Sarektjåkko Stortopp, reaches 2,090 m, while Laitaure is situated 494 m above sea level. Upstream of Litnok the drainage basin has a mean elevation of 1,220 m (*Melin*, 1954). Upstream of the Laitaure delta its mean elevation therefore ought to be about 1,200 m.

The mountains northwest of Laitaure belong to the eastern part of the Caledonian mountain

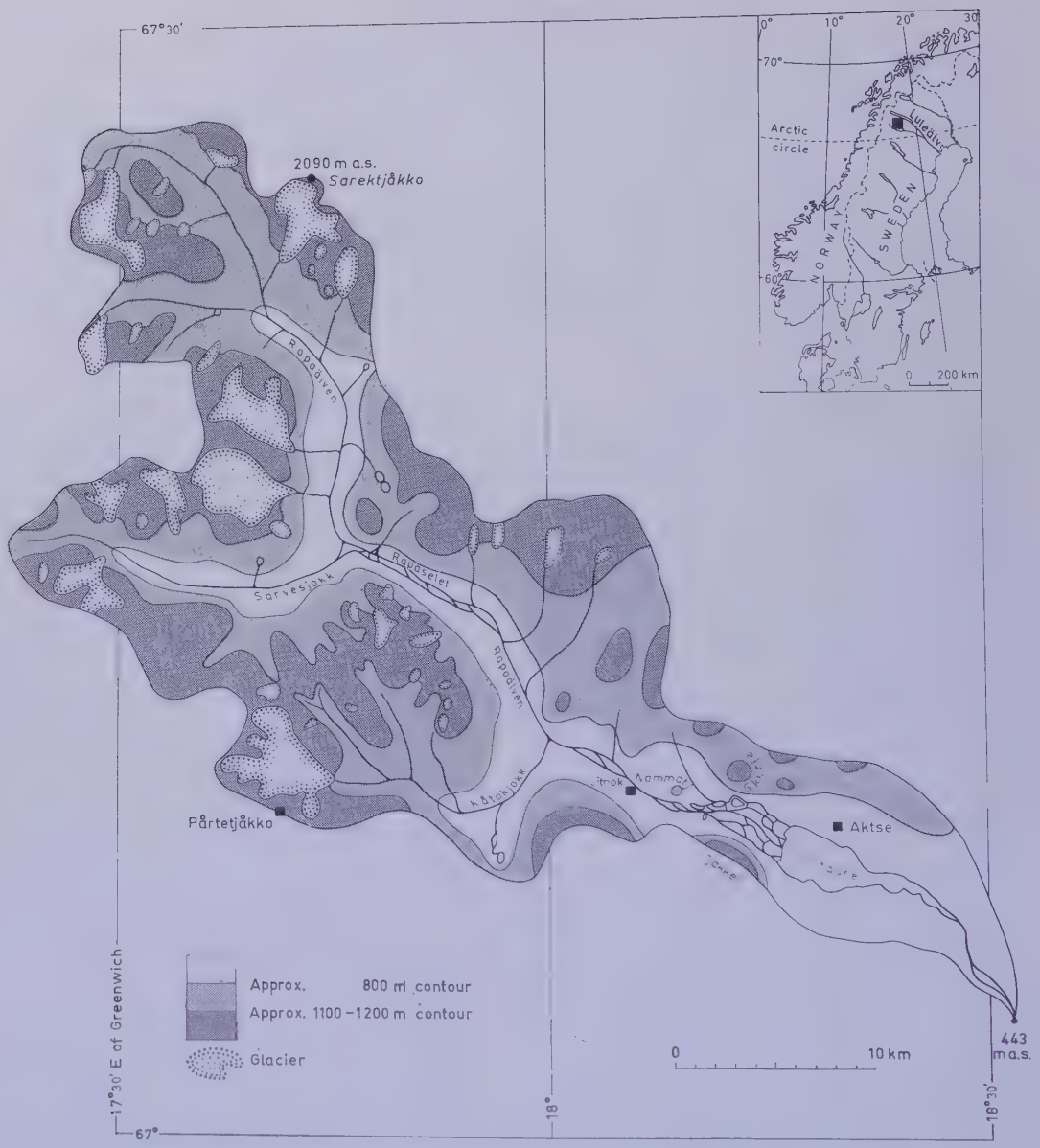


Fig. 21. The drainage basin of the river Rapaälven.

range, and consist mainly of overthrust rock-masses, which during the Caledonian diastrophism were transported from the geosyncline area towards the southeast. *Hamberg* (1910 a, 1915 a) divided these thrust rocks into an upper thrust unit, the Amphibolite Nappe, and a lower thrust unit, the Syenite Nappe. The Amphibolite Nappe, belonging to the lower part of the Seve-köli Nappe (*Kulling*, 1960; 1964), consists mainly of amphibolites. Quartzites and garnet-mica schists also characterize large parts of this nappe. The higher parts of the Sarek mountains consist of rocks belonging to the Amphibolite Nappe. Large areas in the middle part of the drainage basin are built up of rocks belonging to the Syenite Nappe. This nappe consists of far-transported basement rocks,

Fig. 22. View from Lake Laitaure looking upstream. The mountains bordering the delta are: Skerfe to the right; Tjakkeli to the left; and the lower Nammatj in center background. August 11, 1959. Water stage: 103 cm.

mainly schistose syenite and granite. Directly on the Pre-Cambrian basement rests a sequence of autochthonous sediments, the Hyolithus series, mainly consisting of shales and sandstones. The name of this sedimentary series is taken from the fossil *Hyolithes sp.* of Lower Cambrian age, which occurs in these sediments (*Svenonius*, 1884; 1900). In the lower part of the valley of Rapaälven this series appears as a narrow zone along the Caledonian mountain slopes, often almost covered by debris from the above-lying thrust-rocks, exposed in the scarps. Tillite rocks of Eocambrian age, representative of the Varanger ice age and resting on the Pre-Cambrian basement, are observed in the lower part of the valley of Rapaälven (*Kulling*, 1951). In this area the Pre-Cambrian basement of the Caledonian mountains consists of early orogenic Svionian intrusives, mainly of granitic composition. Age determinations indicate that the Svionian cycle terminated about 1,800 million years ago.

The southeastern part of the drainage basin consists of volcanic and sedimentary, Pre-Cambrian rocks, belonging to the oldest members of the Svionian cycle in northern Sweden, the Kiruna-Arvidsjaur complex. In this area the tectonic lines run chiefly in the direction north-northeast, and the structure is developed as a large monoclinal fold or flexure. Acid lavas with beds of basic lavas, tuffs and limestone, dipping to the east-southeast, are at the eastern part of Laitaure conformably followed by the Snavva sediments, a series consisting of feldspathic quartzites interbedded with conglomerates and basic lavas (*Ödman*, 1947; 1957).

During the final phase of the last glaciation the Sarek region was a centre of marked glacial activity (*Hoppe and Liljequist*, 1956; *Hoppe*, 1960).

In the lower part of the valley of Rapaälven the direction of the younger glacial striae mainly follows the direction of the valley. Series of one-sided, lateral "meltwater channels" are found, among others, on the mountain slope north of Laitaure, where they slope to the southeast (*Elofsson and Jansson*, 1961). More or less dissected glacifluvial accumulations, partly built up against damming ice, are found at the lower, hanging part of several tributary valleys. Frost-split rock waste caps large parts of the higher areas, but at lower levels the loose deposits consist predominantly of till, often sandy and often rich in boulders.

Conifer forests are found only in the southeastern part of the drainage basin. In this area the pine occurs only in the lower parts of the prealpine conifer belt. The upper limit of the continuous conifer belt, formed by the spruce, falls considerably towards the west; from an elevation of about 675 m at the eastern end of Laitaure to an elevation of about 500 m at the western end of the lake. Birchwood, however, is found rather far into the larger valleys. In this area the upper limit of the birchwood does not fall towards the west. On the south slopes of the valley-sides the limit between the birchwood and the woodless alpine belt reaches an elevation of about 800 m (*Tengwall*, 1920).

Climate

The northwestern part of the drainage basin is more easily accessible to maritime air masses from the Atlantic than the southeastern part. The climate therefore changes gradually from a maritime to a more continental type towards the southeast (cf. *Atlas över Sverige*, 1953, 1965). The climatic elements, however, vary markedly with altitude and exposure, and the drainage basin is therefore principally characterized by a great variety of local climates.

The meteorological observations in the high mountain region may be considered as started in the summer of 1895, when *Axel Hamberg* placed a minimum thermometer on the highest peak of Sarektjåkko (2,090 m above sea level). Later various meteorological elements have been recorded at several localities within and at the border of the drainage basin, but only at three localities, Pårtetjåkko, Aktse, and Litnok, the continuous records cover a period of several years. For location, see Fig. 21.

At the Pårtetjåkko observatory (1,834 m above sea level) continuous meteorological observations were made during the period 1/7 1914—15/9 1918. A description of the observatory is given by *Hamberg* (1931). At the meteorological station of Aktse (525 m above sea level) the observations only include the precipitation and the air temperature, which from the hydrological point of view are the most important meteorological elements. The observations of precipitation were started in 1937, those of temperature in 1956. At Litnok (505 m above sea level) the continuous records of various meteorological elements as well as of the water level in Rapaälven were begun in 1915. The meteorological records, however, are only partly worked out. The Litnok hydrometeorograph is briefly described in the hydrological chapter.

Air temperature

The annual variation of temperature (°C) is illustrated in Fig. 23 and 24 by means of monthly mean values for Pårtetjåkko and Aktse. The values for Pårtetjåkko are given by *Hamberg* (1932, p. 104). The values for Aktse are compiled from *data* published by *SMHI* (1963—1966) and from diaries kept at *SMHI* (the Swedish Meteorological and Hydrological Institute).

The temperature conditions at Pårtetjåkko and Aktse may be regarded as fairly representative of the highest and lowest parts of the drainage basin. The periods of observation compiled in Fig. 23 and 24 are, however, too short to give reliable mean values. At Pårtetjåkko the mean annual temperature was −8.2° during the period 1/7 1914—30/8 1918, and at Aktse −1.0° during the period 1957—1961. At the meteorological station of Kvikkjokk (337 m above sea level), situated 22 km south of Pårtetjåkko, the mean annual temperature was −1.9° during the period 1/7 1914—30/8 1918 (*Hamberg*, 1932, p. 104); −0.3° during the period 1957—1961; −1.0° during the standard period 1901—1930; and −0.6° during the standard period 1931—1960 (according to *data* published by *SMHI*, 1963—1966). Thus at Kvikkjokk and probably also in Sarek the period 1/7 1914—30/8 1918 was colder, the period 1957—1961 somewhat warmer than the standard periods.

At Aktse the mean annual range (the difference between the means of the warmest and coldest months) amounted to 27.1° during the period

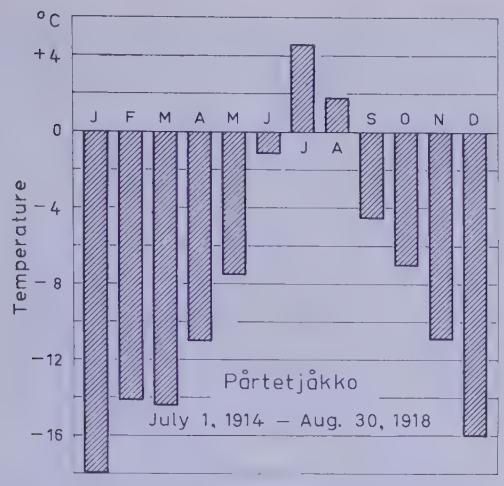


Fig. 23. Monthly mean temperature at Pårtetjåkko.

1957—1961. The mean diurnal range was greatest in March (11.6°) and smallest in November (6.8°). For the warmest month (July) the mean maximum temperature was +17.6°. For the coldest month (January) the mean minimum temperature was −17.7°.

According to *Hamberg* (1915 b, 1932) the average decrease of temperature with height is about 0.4°—0.5° per 100 m in Sarek. As a consequence of the occurrence of strongly developed inversions, the rate of decrease is lower in winter (about 0.2°—0.3° per 100 m) than in summer (about 0.6°—0.8° per 100 m). In midwinter the lowest minima are often found in the valley bottoms. In the winters of 1895—1900 the extreme minimum temperatures were about 10° lower at Kvikkjokk than at the highest peak of Sarektjåkko (*Hamberg*, 1901), and in the winters of 1914—1916 about 4° lower at Kvikkjokk than at Pårtetjåkko (according to *data* published by *Meteorologiska Centralanstalten*, 1914—1916; *Hamberg and Jönsson*, 1933; and *Köhler*, 1939). The lowest recorded temperature amounted to −28.0° at Sarektjåkko and to −29.2° at Pårtetjåkko.

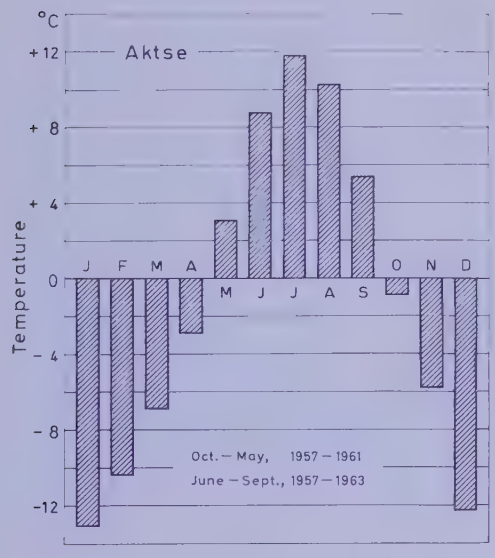


Fig. 24. Monthly mean temperature at Aktse.

At Litnok the mean annual extremes amounted to −40.4° and +24.5° respectively during the period 1952—1966. Temperatures lower than −40° and higher than +24° were recorded 2 out of 3 years. The absolute minimum (− 46.0°) and the absolute maximum (+28.5°) were recorded in 1966.

Lake Laitaure is generally covered by ice from the end of October to the beginning of June (personal communication by *W. Läntha*, Aktse).

Precipitation

The northwestern part of the drainage basin belongs to the wettest part of Sweden with an annual precipitation probably of the order of 2,000 mm in the higher areas. Owing to topographical conditions the amount of precipitation is considerably lower in the southeastern part of the drainage basin. At Áktse the normal annual precipitation amounts to about 600 mm.

The average annual precipitation within the catchment area upstream of Litnok was calculated by *Melin* (1954) on the basis of run-off *data* for the period 1931—1945. However, the value given by *Melin* (1,500 mm) is not adjusted for storage changes and is therefore probably somewhat too high.

For the period 1916—1960 the mean water discharge in Rapaälven was 27.4 m^3/sec. at the proximal part of the Laitaure delta (corresponding to 864 million m^3/year, or 40.1 litres/sec. and km^2). The evaporation within this area amounts to about 150 mm/year (*Melin*, 1942; 1954). According to *Norling* (1957, p. 78) the Mikka glacier had an average annual net loss of volume corresponding to about 0.7 million m^3 of water per km^2 during the period 1913—1954. If we assume that the value for the Mikka glacier is valid also for the other glaciers within the drain-

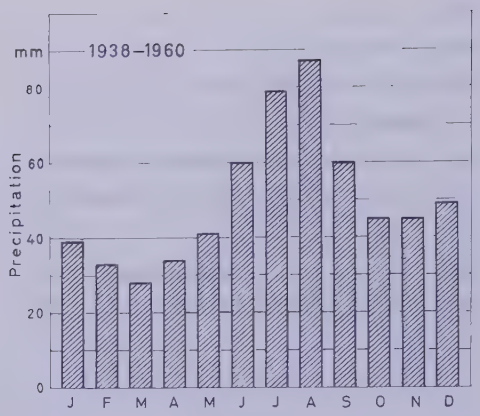


Fig. 25. Monthly mean precipitation at Aktse.

age basin, the average annual net loss of storage volume would have corresponded to about 56 million m^3 of water. A calculation based on the above given figures for run-off, evaporation, and storage changes gives a value of about 1,330 mm for the average annual precipitation within the catchment area upstream of the Laitaure delta. Probably more than half of this precipitation falls as snow.

The published precipitation *data* from Pårtetjåkko comprise only the period 1/7 1914—1/7 1916 (*Hamberg and Jönsson*, 1933; *Köhler*, 1939). During this period the mean precipitation at Kvikkjokk was about 53 % of that at Pårtetjåkko (according to *data* published by *Meteorologiska Centralanstalten*, 1914—1916). If we assume this relation to be valid also for the period 1931—1960, the normal annual precipitation would be about 1,100 mm at Pårtetjåkko.

At Aktse the mean annual precipitation amounted to 599 mm during the period 1938—1960 (according to *data* published by *SMHI*, 1939—1962). The highest annual precipitation was 818 mm, the lowest 437 mm. The monthly mean precipitation at Aktse during the given period is illustrated in Fig. 25. August was normally the wettest month, March the dryest. In winter December showed the highest precipitation. About 40 % of the annual precipitation fell during the winter half-year (October—March).

Deltaic area

Several lakes which formerly existed upstream of the Laitaure delta have been filled up by delta sediments. The largest of these deltaic formations, Rapaselet, has a length of 10 km and is situated about 20 km upstream of the Laitaure delta at an elevation of about 590 m (for location, see Fig. 21).

The only remaining larger lake is Laitaure. This lake formerly had a length of 15 km but is nowadays half filled up by a delta. This delta, the Laitaure delta, has an area of only 10 km^2 but is nevertheless one of the largest and most rapidly growing lake deltas in Sweden. The field works carried out by the author on this delta were started in 1953 and ended in 1959, and are partly presented in some preliminary reports (*Axelsson*, 1955; 1956; 1957; 1960; 1963).

Physiographically this delta has in the main the normal appearance with numerous active and abandoned river channels, more or less marked levee ridges, and shallow inter-levee basins of different types, as shown by the photographs (Fig. 26 and 27). The shape of the delta plain, however, is not lobate but elongate, because its lateral extent is limited by the small width of the original lake basin. Also the splitting and closing of channels as well as the formation of central and marginal delta basins are influenced by initial topographic irregularities of the relatively shallow frontal lake basin.

The maps of the Laitaure delta and Lake Laitaure (Plates I and II) are mainly based on air photographs (BDu 54 01:04—10) taken on June, 18, 1954, from a height of about 3,900 m and having a negative scale of about 1:20,000. The photogrammetric surveys were performed in 1962 with a Wild A5 autograph by *Rikets allmänna kartverk* (the Geographical Survey Office of Sweden). The maps were originally drawn on a scale of 1:4,000. According to *Rikets allmänna kartverk* these original maps have a mean error of about ± 2 m in plan (sharp details) and ± 1 m in height (open terrain). The scale was determined from base-lines and height differences measured by the author.

For the sake of simplicity the contour lines on Plates I and II are given in metres above and below the water stage that prevailed in Laitaure when the above mentioned air photographs were taken. According to calculations by the author and to observations by *W. Läntha*, Aktse, this water stage amounted to 100 cm ± 5 cm above a locally selected level. The water stage 100 cm represents low water in summer and corresponds to an elevation of 494.3 m above sea level. The duration of water stages in Laitaure higher than

Fig. 26. The Laitaure delta. Air photograph by *Rikets allmänna kartverk*, June 18, 1954. Water discharge (Q_R): 38 m³/sec. Approximate scale 1:30,000.

Fig. 27. View from Nammatj looking downstream. The narrow outlet from Lake Laitaure is shown in center background. July 27, 1959. Water discharge (Q_R): 76 m³/sec.

this value amounts on average to about 90 days per year, and is illustrated in Fig. 35 for the years 1957 and 1958.

At the proximal part of the delta narrow parts of the highest levee crests reach somewhat more than 2 m above the given elevation for Laitaure (about 2.1 m above Laitaure at section a—b on Plate I). It was not possible, however, to illustrate this or other height differences on the delta plain by contour lines of desirable accuracy on the basis of photogrammetric surveying. It was therefore decided that the height differences should be illustrated by subdividing the delta plain into areas with and without bushes and trees, indicating higher and lower ground respectively, and, that the delta basins should be subdivided into areas where the prevailing water surface was lower and higher respectively than the occurring vegetation. Plate I may therefore be regarded as a map showing relative height differences as regards areas capped by delta sediments.

Lake Laitaure

The actual elevations of local bench marks established in 1957 near the gauge station on the northwestern shore of Laitaure (see Plate II) were determined in 1959 by assuming the still-water level to be horizontal along the northwestern shore, and by levellings along the northeastern shore to a bench mark of known actual elevation established at the outlet from Laitaure by *Kungl. Vattenfallsstyrelsen* (the State Power Board). It was then found that the figure for the water-surface elevation of Laitaure (498 m) given by the topographic map (13 St. Sjöfallet) was about 4 m too high.

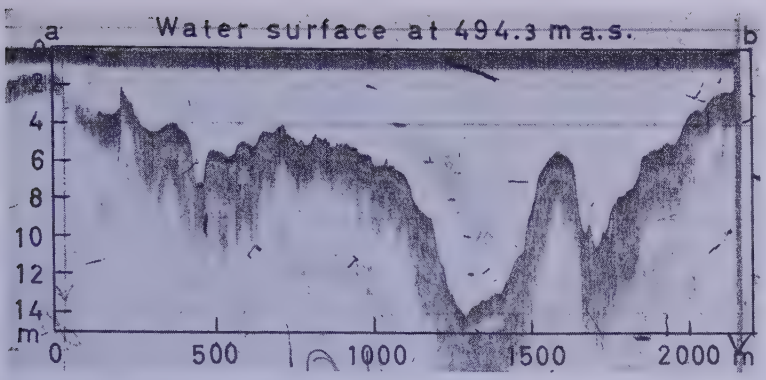


Fig. 28. Echogram of transverse bottom profile from the middle part of Lake Laitaure. Vertical exaggeration × 59. For location of this profile (a—b), see Plate II.

On Plate II the shore-contour of the delta at the given water level is based on surveyings by *Widersten* in 1958. The depth contours are based on soundings by hand (mainly within near-shore areas) and on soundings with a recording echo-sounder (Atlas Monograph), working on 80 MHz.

Within the near-shore areas the sounding positions were determined by measuring the distance along selected directions with a graduated wire, stretched above the water surface. The depths were determined with a measuring lath down to 2.5 m and with a sounding lead at greater depths. Positions far from the shore were determined by angle measurements (with a surveying instrument, Wild N10, having an azimuth scale; with an angle prism; and with a sextant). Echo soundings were made only in calm weather. During the echo soundings the transducer was placed in a special rig at the side of the boat, and at 0.2 m below the still-water level. The boat was steered towards natural or artificial marks in line with each other, and the passages of buoys of known position were marked in the echogram by short-circuiting the pen circuit. 82 profiles, having a total length of 64 km, were sounded downstream of the detailed mapped area (Map C, Fig. 89).

As shown by the hydrographic map (Plate II) and exemplified in more detail by the echogram (Fig. 28) the bottom configuration is very irregular. The greatest depths are found in the central part of the rectangular, western lake basin. The narrow, shallow and boulder-rich outlet bay, Kuolleluoppal, which forms a transition between the lake and the outlet river, is not shown on Plate II. Laitaure has a maximum depth of 17.8 m, a mean depth of 4.1 m, and a mean depth of 4.3 m upstream of section c—d on Plate II. As shown by the morphometric diagrams (Fig. 29 and 30) the lake area (exclusive of islands) amounts to 10.0 km^2, and the water volume to 41.4 million m^3. Upstream of section c—d the lake area amounts to 9.5 km^2, and the water volume to 40.9 million m^3.

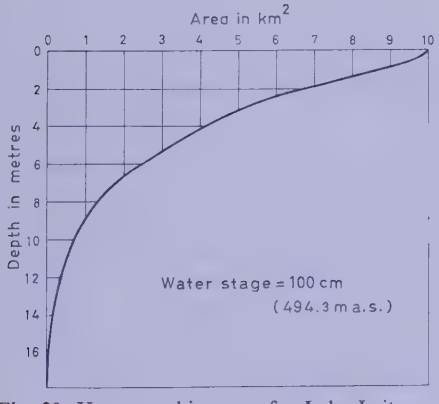


Fig. 29. Hypsographic curve for Lake Laitaure.

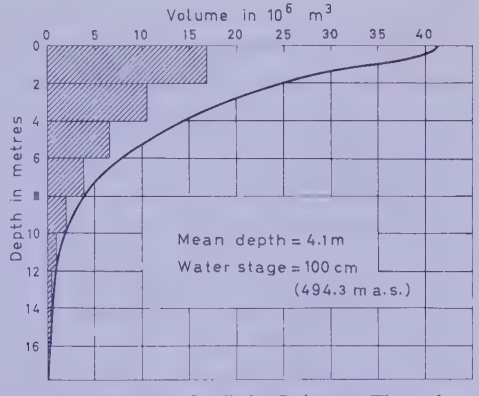


Fig. 30. Volume curve for Lake Laitaure. The volume between given contours is shown by striated piles.

Chapter 8

HYDROLOGY

Gauge stations

The first observations of the water-stage variations in Rapaälven, covering the period 1/8 1897—31/10 1900, were carried out at the proximal part of the Laitaure delta (just downstream of section a — b on Plate I). Three discharge measurements were also performed during this period, an approximate rating curve was constructed, and daily run-off *data* were calculated (*Hamberg*, 1899; 1901; 1930). The water-stage observations, however, were made only once a month in winter and about twice a month in summer. The number of water-stage observations as well as the number of discharge measurements was therefore too small to give satisfactory information on the annual variation of water stage and water discharge.

The Litnok hydro-meteorograph

In order to obtain continuous *data* of the water stage in Rapaälven and also of certain meteorological elements *Hamberg* mounted a specially designed hydro-meteorograph at a fast-flowing part of the river, just below the upper rapids at Litnok (Fig. 31). This instrument was brought into use in the summer of 1915 and, with certain interruptions, has been in function ever since. A description of the instrument was published in German by *Ernfors* in 1938, and in English by *Melin* in 1954. *Melin* (1954, pp. 139—140) described the function of the instrument as follows:

> "A paper moves over a table. Pins connected to instruments for measurement of temperature, humidity, wind, precipitation, and water level, are struck into the paper by a hammer at half hour intervals so that curves are formed by each set of pin holes. The work is done by a pendulum clock with a very heavy weight, which falls some metres a year. The clock runs for 15 months without being wound up, but the paper is changed once a year.....
>
> The water level in the river is transferred to the pin through a float swimming in a vertical steel pipe in the river bank, which is connected with the water by a horizontal pipe. To prevent the water from freezing in the gauge pipe, the latter is filled with 10 litres of oil with a specific gravity lower than the water, and to prevent the oil from escaping into the river at low water a tank is welded to the lower end of the pipe, but above the point from which the horizontal pipe enters the river.
>
> The oil column is constant at high water level when its lower surface is above or just level with the top of the tank. When the water level falls, oil is gathered in the tank and the height of the oil column is diminished as long as there is some oil in the pipe above the tank. Thus the curve on the paper is a function not only of the water level but also of the height of the oil column, and variable corrections must be applied to the oil level when calculating the water level.
>
> As the Rapaälven drains many glaciers, its water in summer-time is very turbid and the horizontal communication pipe may become silted up. When this risk exists a water tank on the top of the instrument hut can be coupled to the gauge pipe and water under pressure be washed through the conduit."

The zero datum of the oil and water stages is set at the bottom of the tank, situated 560 cm below the upper edge of the vertical gauge pipe. Levellings from local bench marks (established just upstream of the station) prove that this zero datum is constant. The volume and the density of the oil has been measured in connection with these levellings. The last check was made in 1957.

The relation between the oil stage above this zero datum (h_o) and the distance of the records from a "northern" zero line on the paper (x_n) is given by the equation

$$h_o = C(x_n - l) \qquad (11)$$

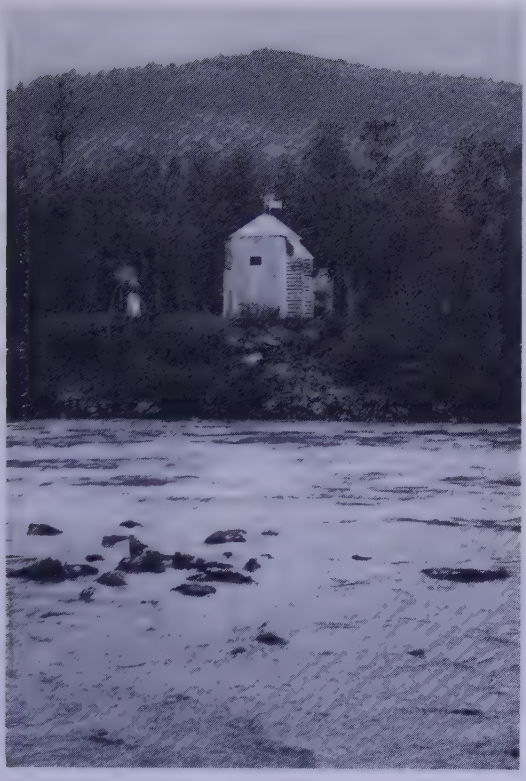


Fig. 31. The hydro-meteorograph hut at Litnok. August 8, 1958. Water discharge: 50 m³/sec.

where the constant C has a value of 14.44, and l is the distance between the "northern" zero line and the true zero line of the records. Since the value of l is not constant but has generally increased with time, it must be calculated. These calculations are based on direct measurements of the distance between the oil level and the upper edge of the gauge pipe. These measurements as well as tests for checking the time-scale and the temperature records are performed each time the station is visited (at present four times a year). During the 1950's the time-scale given by the instrument had to be reduced by about 30 minutes (29—32 minutes) every fortnight.

Oil stage *data* have been compiled by *SMHI* for the period 1915—1945, and by the author for the period 1946—1960. For calculations of water stages the author made use of the relation between oil stage and water stage given by *Melin* (1954, p. 144).

The air temperature at this station is recorded by two thermographs. Parts of the records given by the "southern" thermograph are worked out to illustrate the daily variation of air temperature. The relation between the air temperature in degrees Celsius given by the "southern" thermograph (T_s) and the distance of the records from a "southern" zero line on the paper (x_s) was given for the tested period (1951—1959) by the equation

$$T_s = 41.3 - \frac{x_s}{C} \qquad (12)$$

When giving x_s in mm the constant C has a value of 1.444.

The Laitaure gauge

The systematic observations of the water stage in Laitaure were started in the summer of 1957 when a sectional staff gauge was established at a small landing-stage on the northwestern shore of the lake (for location, see Plate II). For practical and economical reasons this was the only place where it was possible to obtain frequent observations also during periods not covered by the author's observations. The gauge station is further described by *Zackrisson* (1957). Levellings carried out several times in the summers of 1957—1959 gave the result that the elevation of this staff gauge was practically constant in summer, but that it was changed somewhat in winter.

The calculations of water stages and water discharges in Laitaure are mainly based on water-stage observations at this gauge station.

Temporary gauge stations

During the field works water-stage observations were carried out also at several other localities. Some of these temporary gauge stations are marked on Plate I. During shorter periods water-stage recorders were used at the gauge stations marked on Plate I and at the outlet from Laitaure (about 200 m downstream of section c—d on Plate II).

The water-stage observations at the outlet from Laitaure gave, among others, the result that the water stage at the "Laitaure gauge" was not affected by wind during the periods when water-discharge measurements were carried out at the outlet (at section c—d).

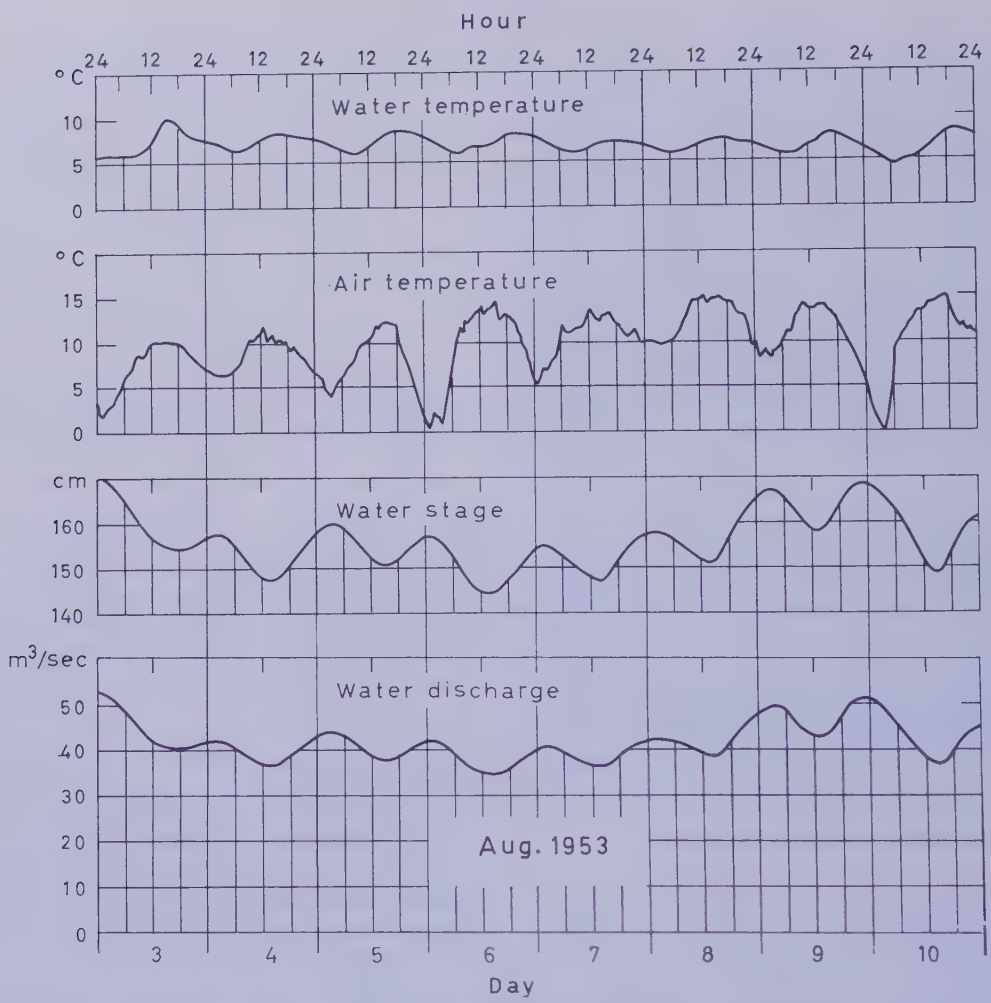


Fig. 32. Water stage, water discharge (calculated), air temperature and water temperature at Litnok during August 3—10, 1953.

Water stage

Rapaälven at Litnok

The annual variation of the water level in Rapaälven is normally of the order of 2.7 m at Litnok. During the period 1916—1960 it varied between 2.2 m (1916) and 3.4 m (1954). The mean annual extreme water stages amounted to 307 cm and 39 cm respectively. The highest water stage probably occurred on August 13, 1954. The very peak was not recorded but was estimated to about 376 cm. The highest recorded water stage amounted to 368 cm (July 3, 1939), the lowest recorded to 32 cm (February 17, 1926).

Due to the absence of larger lakes, the steep topography and the small thickness of the soil cover within the drainage basin, the water level rises and falls very rapidly in summer with changes in precipitation and temperature. The rates of both rise and fall may exceed 1 m in 24 hours.

During dry periods in summer with a marked daily variation in air temperature, there is also a very distinct daily variation in water stage, as well as in water discharge and in water temperature, as exemplified in Fig. 32 for the period

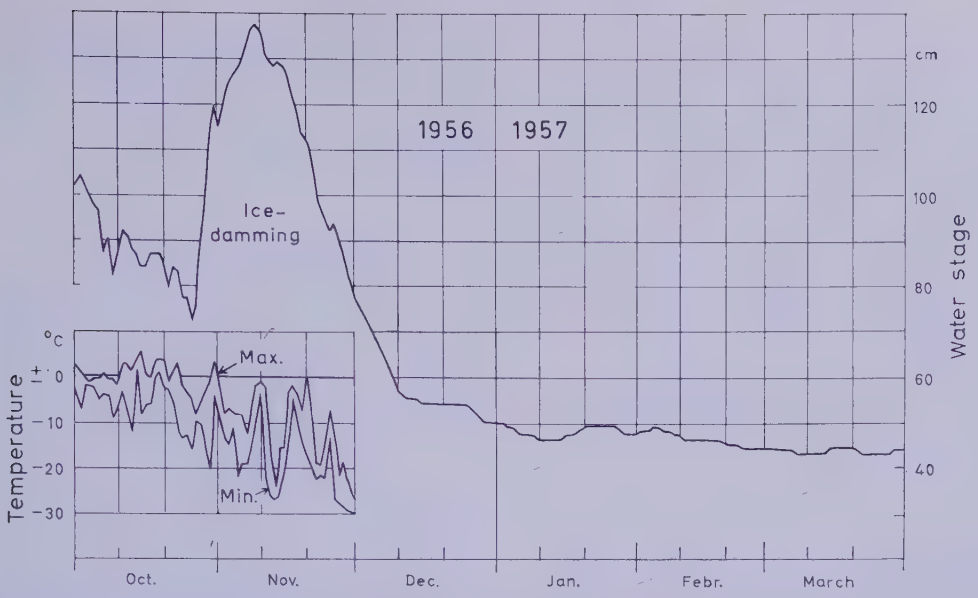


Fig. 33. Daily mean water stage in Rapaälven at Litnok during the winter half-year 1956/57. Daily max. and min. temperature at Litnok during October and November 1956 (determined from half-hour values).

August 3—10, 1953. During this period the time-lag between the daily maximum in air temperature and that in water stage amounted to about 12 hours. The time-lag between the daily maximum in air temperature and that in water temperature amounted to about 3 hours. The daily maximum in air temperature, in water temperature, and in water stage was recorded at about 2 p.m., 5 p.m., and 2 a.m. respectively; the daily minimum in air temperature, in water temperature, and in water stage at about 2 a.m., 7 a.m., and 3 p.m. respectively. The water temperature was recorded by a thermograph, mounted on a specially designed raft and placed in the river just upstream of the instrument hut at Litnok; the accuracy of these records was tested by means of a reversing thermometer (*Axelsson*, 1956). For further examples of the daily variation in water stage at Litnok, see *Melin* (1954) and *Lund* (1959).

As exemplified by Fig. 33, the water level is normally dammed up by the ice in the early part of the winter. This ice-damming usually begins in October with the undercooling of water and

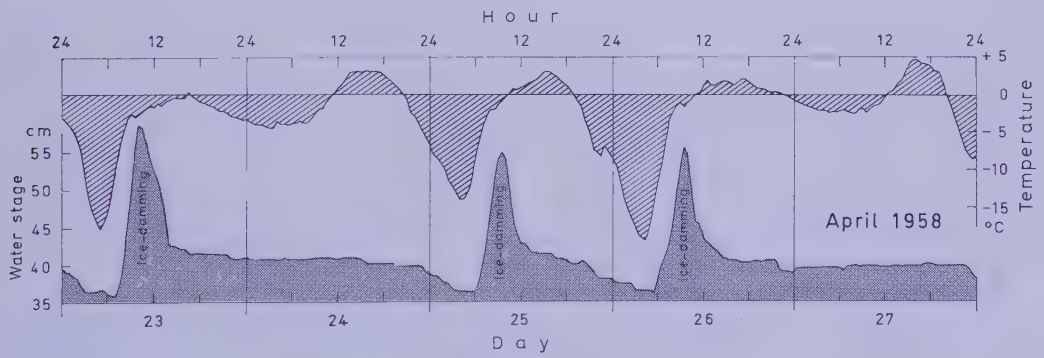


Fig. 34. Water stage and air temperature at Litnok during April 23—27, 1958.

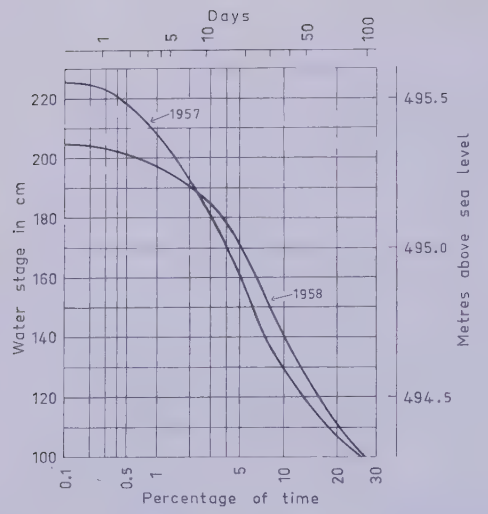


Fig. 35. Duration of water stages in Lake Laitaure higher than 100 cm (494.3 m a.s.) during the years 1957 and 1958.

the formation of frazil and anchor ice. When the rapids are also covered by ice the water level normally falls, leaving the ice as a bridge across the stream, supported by boulders, and in late winter the water level is usually not influenced by the ice.

In April and in the beginning of May, however, a diurnal type of ice-damming often occurs when parts of the rapids at Litnok are ice-free, the diurnal range in air temperature is great, and the nights are very cold. An example of such a situation is illustrated in Fig. 34. During April 23, 25 and 26, 1958 the daily minimum and maximum in water stage was recorded at about 6 a.m. and 10 a.m. respectively; the daily minimum and maximum in air temperature at about 5 a.m. and 16 p.m. respectively.

Laitaure

In Laitaure the normal annual variation of the water level probably amounts to about 1.6 m. Calculations based on observations of water stages in Laitaure during the period 1957—1960 and on the relation between the variation of water stages in Rapaälven at Litnok and in Laitaure give the result that the mean annual extreme water stages in Laitaure (*MHW* and *MLW* in Fig. 50) probably amount to about 210 cm and 50 cm respectively. The mean annual water stage (*MW*) probably amounts to about 90 cm. The duration of water stages higher than 100 cm is illustrated for the years 1957 and 1958 in Fig. 35.

In Laitaure the water level is also normally dammed up by the ice, most markedly in the early part of the winter.

The Laitaure delta

In the proximal part of the delta the annual variation in river stage is normally of the order of 2—2.5 m, and the variations may be rapid as shown by Fig. 36. At the delta front and in the delta lakes the variations are slower. In those delta lakes which are entirely cut off from channel connection with the distributaries the variation in water stage is usually very slow, and the water stage in these lakes may therefore differ considerably from that in the adjacent distributaries. In those delta lakes which have only one channel connection with the active distributaries the connecting channel may function as an inflow during rising river stages and as an outflow during falling river stages. However, if its threshold level is

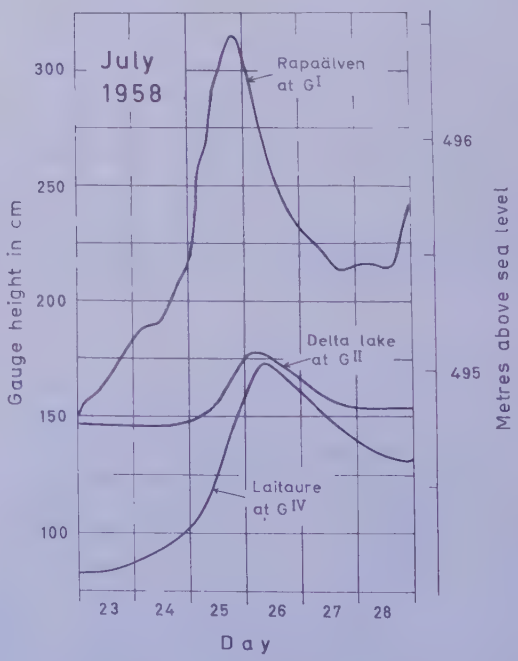


Fig. 36. Water stage in Rapaälven at the proximal part of the Laitaure delta, in a distal delta lake, and in Lake Laitaure at the delta front during July 23—28, 1958. For location of the gauge stations G^{I}, G^{II} and G^{IV}, see Plate I.

Fig. 37. Stage difference between the gauge stations G^{I} (at the proximal part of the Laitaure delta) and G^{IV} (at the delta front in Lake Laitaure) in relation to the water discharge in Rapaälven during July 23—28, 1958.

high it will be active only at high river or lake stages, as exemplified by the water-level diagram for the delta lake at G^{II} (Fig. 36).

The longitudinal slope of the water surface in the deltaic channels varies with the river discharge and with the water stage in front of the channel mouths, and at the delta front the lateral zones of flow separation migrate upstream with increasing storage in Laitaure. The stage difference between gauge I at the proximal part of the delta and gauge IV at the delta front during the period July 23—28, 1958 is illustrated in Fig. 37 in relation to the variation in river discharge. As shown by the diagram the stage difference may be considerably greater during rising than during falling river discharges. This also means that the longitudinal slope of the water surface as well as the flow velocity is greater during rising than during falling river discharge, especially in the lower part of the distributaries. As shown by Fig. 37 the greatest stage difference was reached before the river discharge reached its highest value. Due mainly to higher water stage in Laitaure the stage difference was lower during the second than during the first rise in river discharge. During the given period the greatest and the smallest difference in water-surface elevation between the two gauge stations amounted to 174 cm and 68 cm respectively. These values correspond to an average slope of the water surface along the deltaic channels of about 0.3 and 0.1 per mille respectively.

The younger and lower levees are flooded more frequently and during longer periods than the older and higher levees. The higher levee crests are topped by flood-water only about 2 out of 3 years on the average, even if they are overtopped twice in some years, as, for example, in 1957. The duration of water stages higher than the higher levee crests is very short, and usually less than one day.

According to direct observations in 1954 and in 1957 the bankfull stage at section a—b in the proximal part of the delta corresponded to an oil stage at Litnok of about 323 cm, which represents a water discharge of 280 m^3/sec. on the rating curve (Fig. 43). This value of the bankfull discharge represents on the average the 0.1 % point

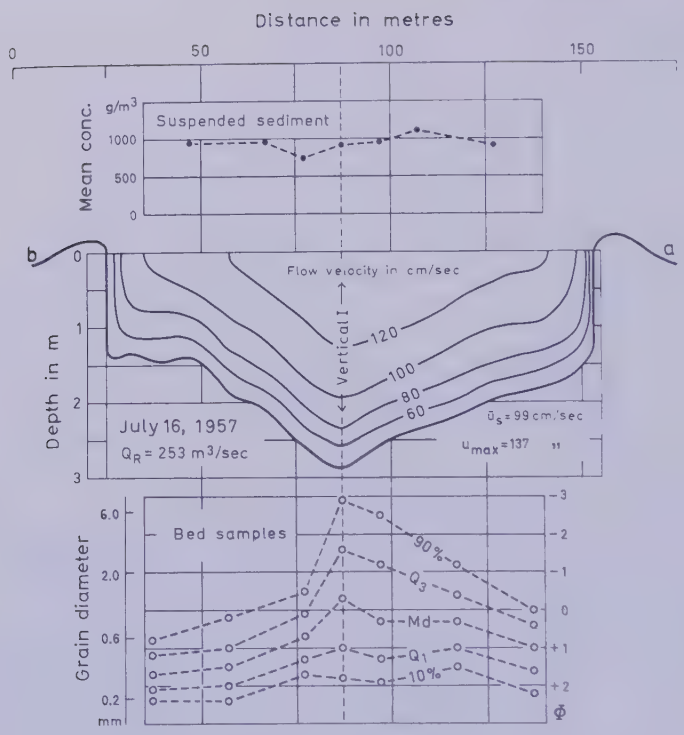


Fig. 38. Velocity distribution in section a—b at the proximal part of the Laitaure delta. Weighted mean sediment concentration in selected verticals. Grain-size distribution of the bed material. July 16, 1957. Vertical exaggeration × 20.

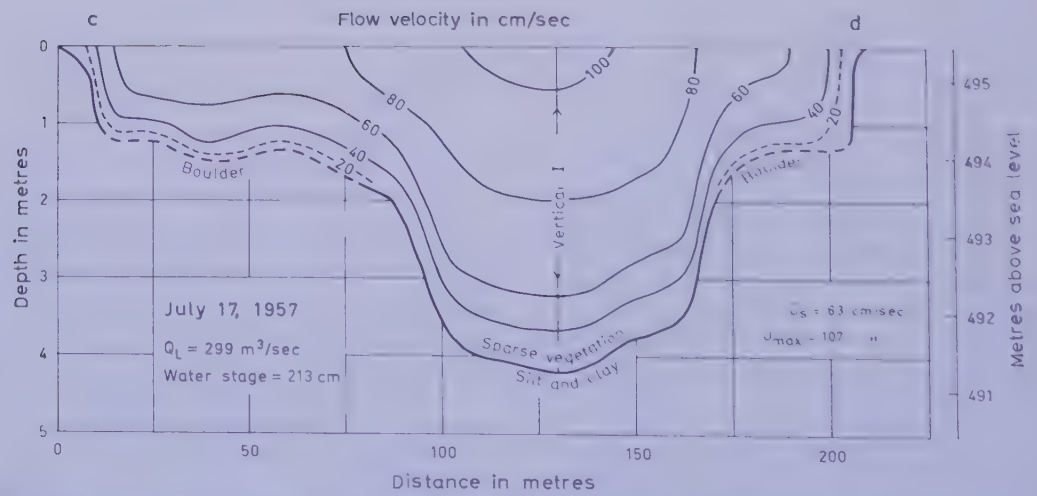


Fig. 39. Velocity distribution in section c—d at the outlet of Lake Laitaure on July 17, 1957. Vertical exaggeration × 20.

on the flow-duration curve (Fig. 48), and an average return period of 1.5—2 years on the flood-frequency curve (Fig. 49).

Flow velocity

Measurements of the flow velocity were made on several occasions in section a—b at the proximal part of the delta and in section c—d at the outlet from Laitaure. The measurements were carried out with an Ott-Kempten current meter (a propeller meter) suspended by a cable.

In the proximal delta section the flow velocities are normally rather symmetrically distributed. The velocity distribution in this section on June 16, 1957 is illustrated in Fig. 38. The velocity diagram is based on measurements in 15 verticals. The distance between successive verticals was 10 m, except close to the river banks, where it was 5 m. On this occasion the water discharge in Rapaälven at this section (Q_R) was 253 m³/sec., somewhat lower than the bankfull discharge. The maximum velocity (u_{max}) was 137 cm/sec. and the mean velocity in the cross section ($\bar{u}_s$) 99 cm/sec. The flow velocities in this section vary with the water discharge and are normally considerably lower at low water. An example of this is that on July 6, 1957 the maximum velocity was 58 cm/sec. and the mean velocity in the cross section 41 cm/sec. when the water discharge was 38.6 m³/sec., which represents low water in summer and approximately the water discharge that prevailed when the previously mentioned air photographs were taken. In this section the flow velocities are somewhat higher during rising than during falling water discharges, but the difference is not as pronounced as further downstream.

Also in the outlet section in Laitaure the maximum velocity may be higher than 1 m/sec. at high water discharges. The velocity distribution in this section on July 17, 1957 is illustrated in Fig. 39. The diagram is based on measurements in 20 verticals. The distance between successive verticals was 10 m. On this occasion the water discharge at this section (Q_L) was 299 m³/sec., which is somewhat higher than the calculated mean annual extreme high-water discharge.

Vertical velocity distribution

The velocity distribution in open-channel flow over a sediment bed is often described by the logarithmic formulas based on *von Kármán's* (1930) similarity theorem with the constants as proposed by *Keulegan* (1938). See, for example, *Einstein* (1950) and *Sundborg* (1956). The logarithmic formulas for smooth flow, transitional flow, and fully rough flow, may be combined in the form:

$$\frac{u}{u_*} = \frac{1}{k} ln \frac{y}{\alpha k_s} + 8.5 = \frac{2.3}{k} \log_{10} \frac{y}{\alpha k_s} + 8.5 \qquad (13)$$

where u is the average point velocity at a distance y above the bed, u_* the shear velocity, k the von Kármán "constant", k_s a length parameter, and α a correction factor, with a value that depends on the ratio of the height of the irregularities projecting from the bed surface (h) to the thickness of the laminar sublayer (δ). The values of α as a function of h/δ are graphically illustrated by *Sundborg* (1956, Fig. 8). For fully rough flow the value of α is 1.

The shear velocity may be written

$$u_* = \sqrt{\frac{\tau_o}{\varrho_f}} = \sqrt{g R S_e} \qquad (14)$$

where τ_o is the boundary shear, ϱ_f the density of the fluid, g the acceleration due to gravity, R the hydraulic radius, and S_e the energy gradient.

The von Kármán "constant" of turbulent exchange (k) was determined by *Nikuradse* (1932, 1933) for pipe flow with clear water. The value obtained was 0.4. Formula (13) is therefore sometimes written

$$\frac{u}{u_*} = 5.75 \log_{10} \left(\frac{30.2\, y}{\alpha k_s} \right) \qquad (15)$$

However, flume and field studies have shown that the value of k may vary considerably. *Sayre and Albertson* (1961) reported k values ranging from 0.34 to 2.77 for clear-water flow with isolated roughness baffles. *Nordin and Dempster* (1963) reported k values ranging from 0.15 to 1.20 in reaches of the Rio Grande and the Rio Puerco. High values of k are usually attributed to the effect of secondary flow. *Rozovskii* (1957) found in his experiments that k in general has a mean value of about 0.5 in bends of open channels, but that the value of k increases along a bend. He considered this as evidence of the gradual evening out of velocities along the vertical under the influence of transverse circulation. Low values of k may be attributed to the effect of high

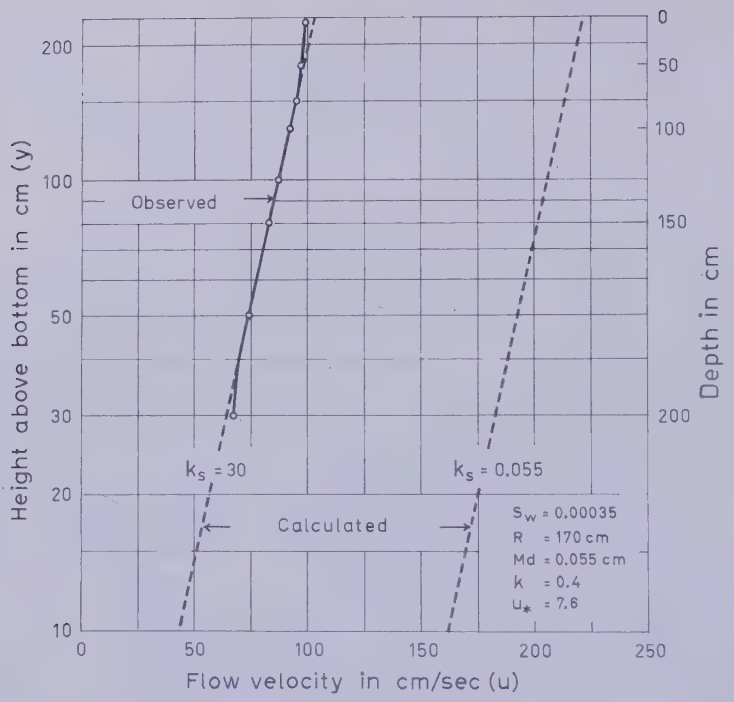


Fig. 40. Calculated and observed velocity distribution in vertical I of the proximal delta section on July 21, 1957.

concentrations of suspended sediment, that dampens the turbulence and reduces the momentum transfer. This effect was studied by *Einstein and Chien* (1955), who measured the reduction of k by the amount of turbulent energy spent in supporting the sediment in suspension. Formula (15) is therefore not valid for reaches where the tendency to secondary flow is pronounced or where the concentration of suspended sediment is high.

The value of k_s in formulas (13) and (15) is dependent on the roughness of the bed surface, and varies therefore with the size and spacing of the roughness elements. According to *Einstein* (1950) the value of k_s for flow over a flat sediment bed equals the grain diameter (d) when the bed consists of uniform material, and the grain size at which 65 per cent of the mixture (by weight) is finer (d_{65}) when the bed consists of non-uniform material. A sediment bed, however, is usually not flat but more or less irregular, and may often show riffles and bars of various shapes and sizes. The value of k_s is therefore usually considerably greater than the grain diameter d_{65}. *Nordin and Dempster* (1963) found in their field studies the mean k_s to be about 400 times as great as d_{65}.

In the proximal delta section, studied by the author, the grain size of the bed material varied along the section and with the flow velocity, as exemplified by Fig. 70. The bed configuration in the adjoining river reach was somewhat irregular, and in the central part normally characterized

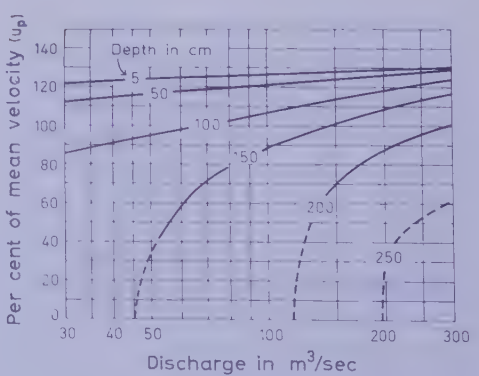


Fig. 41. The relationship between the water discharge in the proximal delta section and the average flow velocity at different depths below the water surface in vertical I of this section in percentage of the mean velocity in the vertical. The diagram is based on measurements in the summers of 1954—1958.

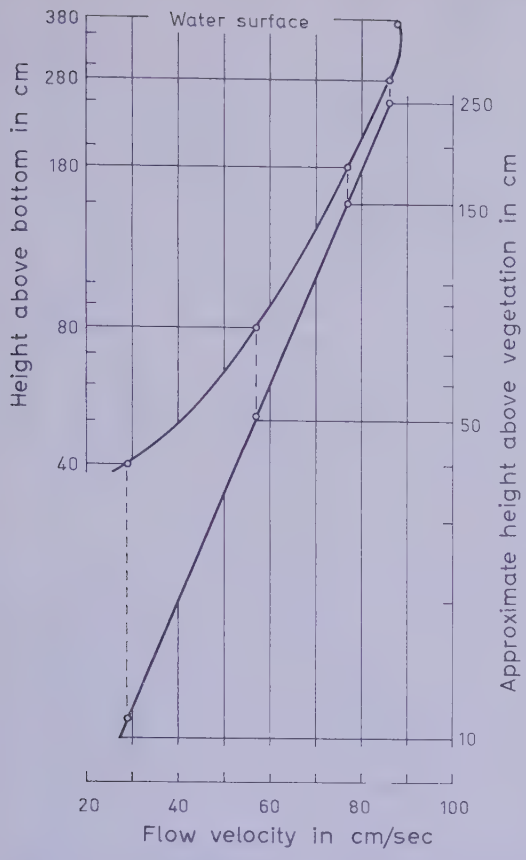


Fig. 42. Velocity distribution in vertical I of the outlet section in Lake Laitaure on July 26, 1957.

This is exemplified in Fig. 40 by calculated and observed velocity distribution in vertical I of the section on July 21, 1957. For location of this vertical, see Fig. 38. On this occasion the value of k_s (30 cm) was about 550 times as great as the *Md* value (0.055 cm) and about 400 times as great as the d_{65} value (0.074 cm) of the bed material in this part of the section. When calculating the shear velocity the energy gradient was taken equal to the slope of the water surface (S_w).

The average flow velocity at different depths below the water surface in vertical I of the proximal delta section is in Fig. 41 given in percentage of the mean velocity in the vertical and in relation to the water discharge in the section. The probable vertical velocity distribution was determined graphically by means of this diagram for those occasions when only samples of suspended sediment were taken at selected points in this vertical.

The sparse vegetation in the deeper part of the outlet section in Laitaure (Fig. 39) reaches a height of about 0.3 m above the bottom at high-water discharges in summer. This vegetation has a damming effect and influences the vertical velocity distribution. The velocity distribution in vertical I of this section on July 26, 1957 is illustrated in Fig. 42. As shown by the diagram the velocity on this occasion was proportional to the logarithm of the approximate height above the vegetation, but not to the logarithm of the height above the bottom.

by the occurrence of transverse bars having a length of about 15 m and a height of the steep distal slope of about 0.3 m. Secondary flow phenomena were observed close to the banks and at the distal slope of the bars. The concentration of suspended sediment was generally not high enough to appreciably affect the turbulent exchange. In this section the flow velocities were normally logarithmically distributed along the vertical, except for regions close to the water surface, close to the river banks, and probably also certain regions close to the bed. The accuracy of the flow measurements, however, was not high enough to give reliable information about the flow velocities close to the bed surface. The mean value of k was about 0.4. The value of k_s varied but was always considerably greater than the median grain diameter (*Md*), or the grain size at which 65 per cent of the bed material is finer (d_{65}).

Water discharge

Rapaälven

For the sake of simplicity the rating curve for Rapaälven (Fig. 43) is related to the oil stage in the gauge pipe at Litnok. The rating curve is based on discharge measurements made by *Melin* in 1935, by *Ernfors* in 1937, and by the author in 1952 — 1958. The measurements made by *Melin* and *Ernfors* were carried out at Litnok. The measurements made by the author were carried out at the proximal delta section, except for the two giving the lowest discharge values, which were carried out at Litnok. The drainage area is about 2 per cent greater at the proximal delta section than at Litnok. Since the precipitation generally is considerably smaller in the lower, eastern than in the higher, western part of the drainage basin the average water discharge ought to be

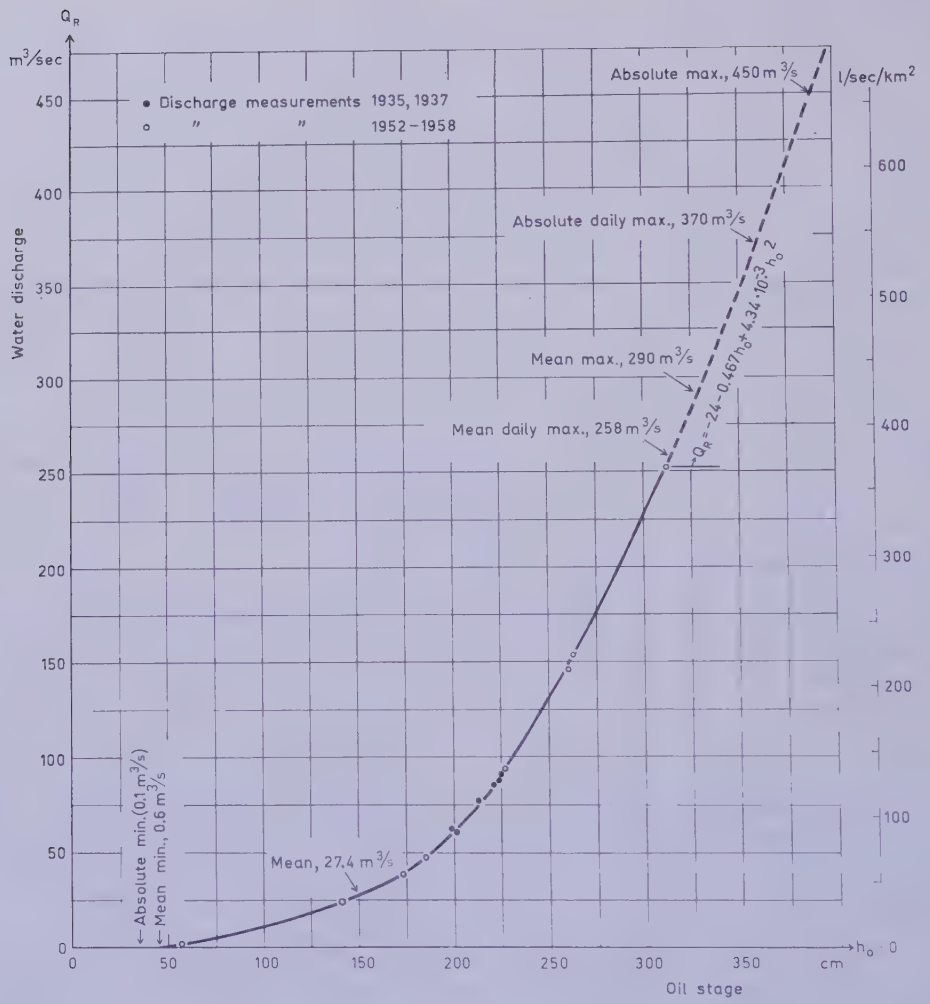


Fig. 43. Rating curve for Rapaälven at Litnok.

only about 1 per cent higher at the proximal delta section than at Litnok. It was therefore not found necessary to correct the measured discharge values with regard to the different location of the measuring sections. The time-lag for the water discharge between the measuring sections is also very insignificant. The lowest measured discharge is 1.7 m³/sec. Beyond the highest measured discharge, 253 m³/sec., the rating curve is represented by means of a parabola, based on the four highest measured discharges and the corresponding oil stages. The equation used for computation of water discharges higher than 253 m³/sec. is

$$Q_R = -24 - 0.467 \cdot h_o + 4.34 \cdot 10^{-3} \cdot h_o^2 \quad (16)$$

where Q_R is the water discharge in m³/sec. in Rapaälven at Litnok or at the proximal delta section, and h_o the oil stage in cm in the gauge pipe at Litnok.

As shown by the rating curve (Fig. 43) the discharge measurements in the 1930's gave the same stage-discharge relation as those in the 1950's. It is therefore probable that the control section is firm and that the rating curve is valid also for the earlier period of oil-stage records at Litnok. The author has therefore used this rating curve for

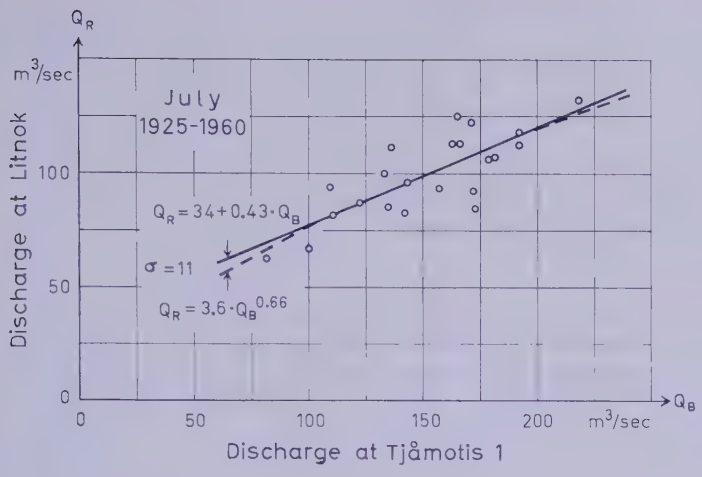


Fig. 44. The relationship between the mean water discharge in Rapaälven at Litnok and the mean water discharge in Blackälven at Tjåmotis 1 for July (1925—1960).

computations of water discharges from the summer of 1915, when the oil-stage records started, to 1960.

The water discharge in Rapaälven at Litnok has been described by *Melin* (1954, 1955) for the period 1931—1945 on the basis of a preliminary rating curve. This preliminary rating curve gave about the same stage-discharge relation as the rating curve used by the author for discharges from 4 m³/sec. to 90 m³/sec., but higher discharge values for discharges lower than 4 m³/sec. and lower discharge values for discharges higher than 90 m³/sec.

For the periods when the water level was influenced by the ice and for the periods when the oil-stage recorder was out of operation the water discharge was computed by interpolation, by comparison with the variation in water discharge at neighbouring gauge stations, and by studies of the meteorological conditions. The water discharge in winter is partly computed by comparison with the variation in water discharge at the outlet of Lake Saggat, situated about 33 km south of Litnok. According to *Melin* (1954) the outlet of this lake is ice-free throughout the winter. For the longer periods when oil-stage records are lacking the monthly mean water discharge has been computed for the months May—September by comparison with the monthly mean water discharge in the river Blackälven at the gauge station Tjåmotis 1, situated about 40 km downstream of Litnok. At this gauge station the drainage area is about 3 times greater than at Litnok. The relationship between the mean water discharge in Rapaälven at Litnok (Q_R) and that in Blackälven at Tjåmotis 1 (Q_B) for July (1925—1960) is shown in Fig. 44. For this month the method of least squares gives the following equation for a straight line approach

$$Q_R = 34 + 0.43 \cdot Q_B \qquad (17)$$

and the following for an exponential curve approach

$$Q_R = 3.6 \cdot Q_B^{0.66} \qquad (18)$$

Both equations give a standard deviation (σ) for Q_R of 11 m³/sec. For the months May, June,

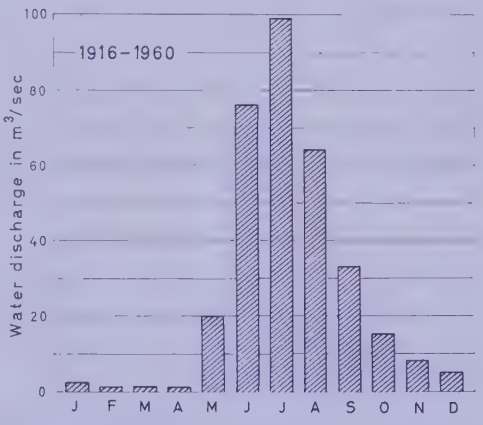


Fig. 45. Monthly mean water discharge in Rapaälven at Litnok (1916—1960).

Year	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Annual mean
1915	-	-	-	-	-	-	-	42	19.4	6	4	0.5	-
16	0.8	1	1	2	13.2	51	81	35	19.9	6	5	1.8	18.2
17	0.8	0.4	0.3	0.3	9.2	106	59	84	26	11.3	6	4	25.7
18	2	2	2	1.6	15.0	61	100	44	25	15	11	6	23.9
19	3	2	0.8	0.6	36	82	80	42	30	10	6	3	24.7
20	0.6	0.4	0.4	0.7	32	78	95	72	34	21	13	6	29.6
21	1.8	0.7	0.7	4.0	35	62	63	76	26	15	8	4	24.9
22	1.5	1.1	0.6	0.9	22	108	95	74	36	9	6	3	29.9
23	1.4	0.7	0.6	0.5	8.0	28	107	55	37	14	7	4	22.1
24	2	1	1	1	9	43	111	90	37	22	10	6	28.0
25	3	2	1	1	15	76	123	67	26	10	5	2	27.8
26	0.9	0.2	0.4	1.0	10.5	75	81	50	29	13	7	4	22.8
27	2	1	1	1	7	102	116	49	31	16	8	5	28.4
28	2	1	1	1	12	60	97	86	31	11	8	6	26.5
29	3.9	1.9	2.3	1.2	23	78	67	61	39	18	9	6	26.0
30	4	2	2	1.7	36	82	98	51	29	13	9	6	28.0
31	2.8	1.4	1.3	0.8	34	31	113	70	19.4	14	8	5	25.4
32	3	3	1.7	0.6	11.5	64	107	64	31	9	6	5	25.6
33	4	1.4	0.7	0.6	7.4	86	111	66	37	13	8	6	28.6
34	2.1	1.5	0.9	0.6	32	69	125	79	55	16.2	10	7	33.4
35	4.1	2.1	2.1	1.2	2.2	107	131	41	27	16	9	6	29.2
36	2.7	1.3	0.8	0.7	30	115	113	84	34	11	8	6	34.0
37	4	1.4	0.8	3.4	38	90	94	72	32	18.9	10	7	31.2
38	4	1.9	1.3	1.5	8.3	119	126	59	29	43	12	8	34.6
39	6	4	1.9	1.4	9.3	65	118	70	40	11	7	5	28.4
40	2.1	1	1	1	45	66	62	70	45	20	8	4	27.2
41	2	1	0.9	0.5	16.4	53	100	86	38	17	9	4	27.6
42	0.9	0.5	0.8	2	22	66	84	56	30	16	10	6	24.7
43	3	2.0	1.0	0.7	12.4	101	106	57	28	13	8	6	28.3
44	4	1.7	0.9	0.5	6.0	64	92	71	32	26	9	7	26.3
45	4	1.4	1	2	22	92	112	74	29	16	8	6	30.8
46	3	2	1	1	14	78	80	64	52	22	10	7	28.0
47	3	1.6	1.2	1.4	18.5	82	86	47	31	11	7	4	24.6
48	0.9	0.5	1	1	26	51	85	56	31	12	8	6	23.3
49	3	2	1	2	34	91	93	56	33	16	9	6	29.0
50	3	2	1	1	14	93	112	60	34	20	10	7	29.9
51	3	2	1	1	7	57	118	90	35	14	8	6	28.8
52	3	2.2	2.5	1.6	14.1	95	120	35	18.4	9	5	1.8	25.7
53	0.8	0.5	0.8	0.8	20	102	75	67	34	17	9	5	27.8
54	3	1.6	1.1	1.0	44	56	93	92	34	13	8	6	29.7
55	3.7	2.3	2	1	4	61	117	53	39	17	9	5	26.3
56	2	1	1	1	13	71	93	52	23	10	6	3	23.1
57	1.4	1.2	1.0	1	11	64	132	64	45	10.9	7	4	28.8
58	2	2	0.7	0.7	7.6	88	82	66	36	16.1	9	5	26.4
59	3	2	1.4	1.7	40	64	85	54	28	31	9	5	27.2
60	2.0	1	1	1	44	73	96	84	37	12	8	4.4	30.4
Mean 1916-60	2.6	1.5	1.1	1.2	19.8	76	99	64	33	15.4	8.2	5.1	27.4

Table 1. Monthly and annual mean water discharge in Rapaälven at Litnok.

August, and September a straight line approach gives a standard deviation for the monthly mean water discharge in Rapaälven at Litnok of 6, 12, 6, and 5 m^3/sec. respectively.

For the 45 years' period 1916—1960 and for the 30 years' period 1916—1945 the mean water discharge was 27.4 m^3/sec., corresponding to an average specific run-off of 40.1 litres/sec. and km^2 within the drainage basin upstream of the Laitaure delta. This value of the run-off is very high in comparison with most drainage basins of comparable size in Sweden. During the period 1916—

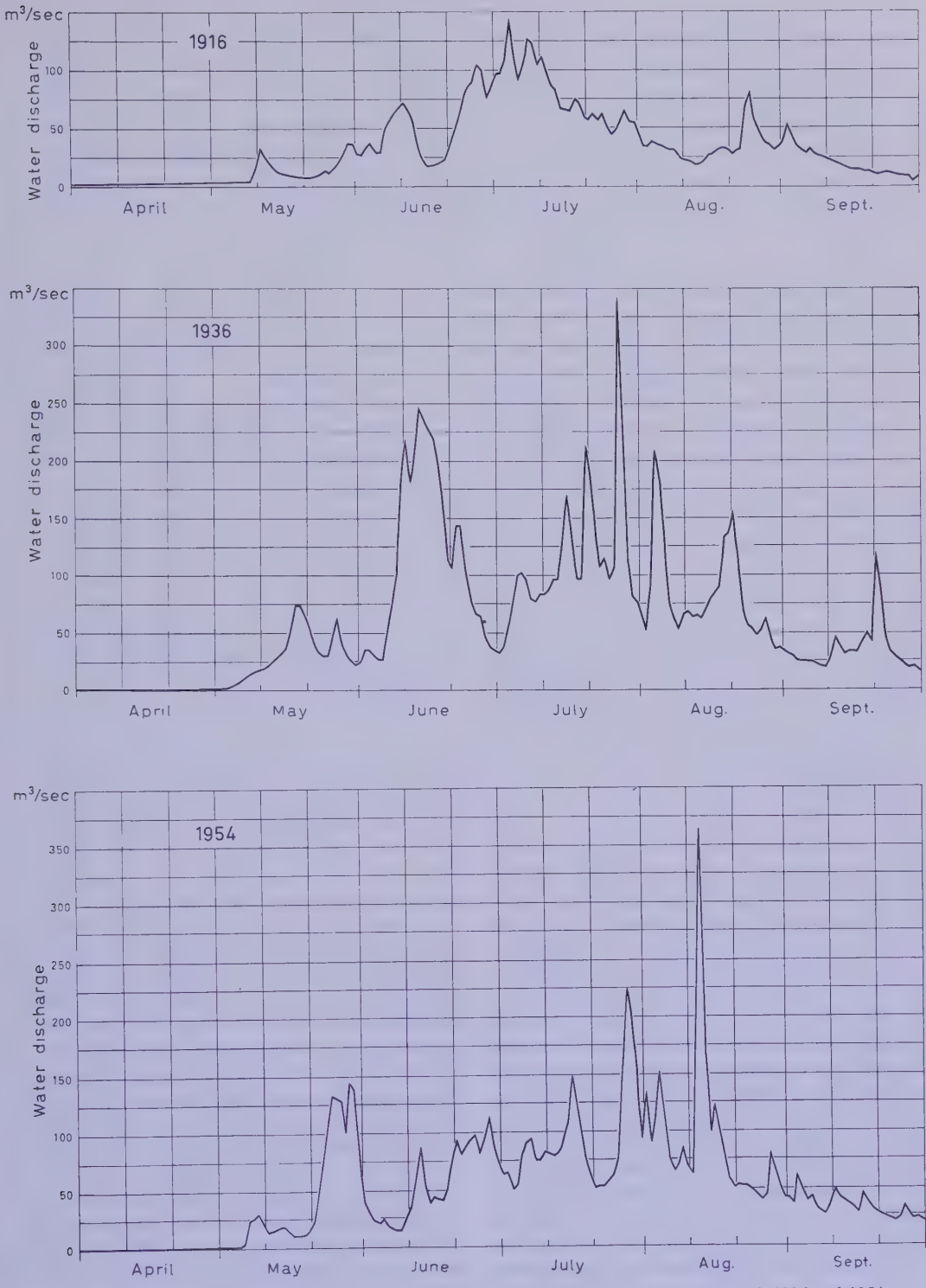


Fig. 46. Daily mean water discharge in Rapaälven at Litnok during the summer half-years 1916, 1936 and 1954.

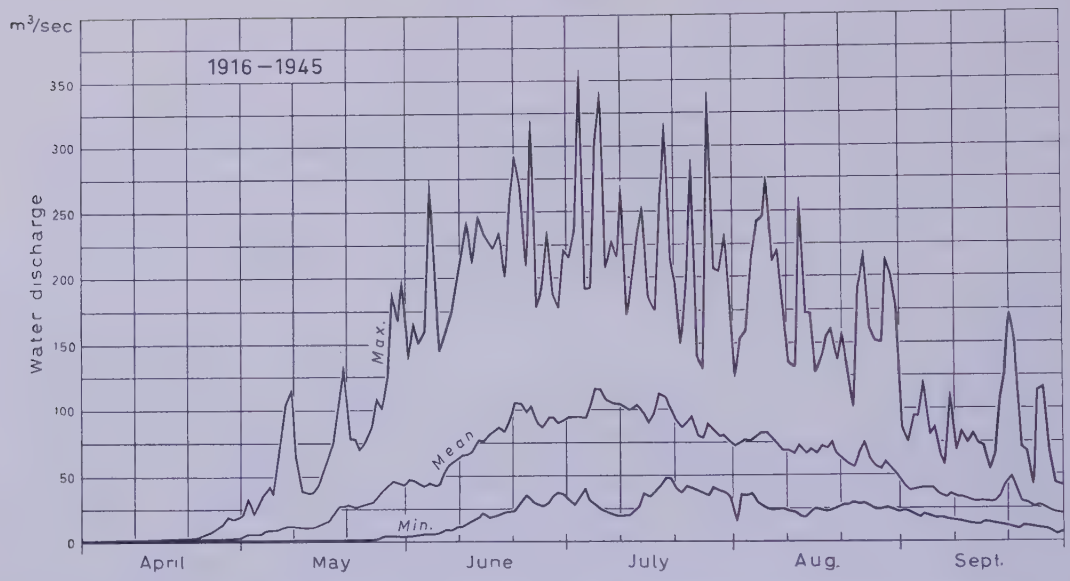


Fig. 47. Mean and extreme water discharge in Rapaälven at Litnok during the summer half-year, determined from mean daily flows (1916—1945).

1960 the mean annual water discharge for the calender years varied between 18.2 m³/sec. (1918) and 34.6 m³/sec. (1938) or between 66 % and 126 % of the mean value for the period (see Table 1). For the water years (October 1 to September 30) it varied between 18.0 m³/sec. (1915—1916) and 33.9 m³/sec. (1935—1936) or between 66 % and 124 % of the mean value.

The annual variation in water discharge is very great. The highest recorded oil-stage corresponds to a water discharge of 452 m³/sec. (July 3, 1939). The peak discharge on August 13, 1954 was probably higher and was estimated to about 475 m³/sec. The lowest recorded oil-stage corresponds to a water discharge of about 0.1 m³/sec. (February 17, 1926). The mean annual extreme values amount to 294 m³/sec. and 0.6 m³/sec. respectively. The absolute and the mean annual maximum flows, determined from mean daily flows, amount to 370 m³/sec. (August 13, 1954) and 258 m³/sec. respectively. These extreme values, especially the absolute extreme values, must be regarded as approximate, since they are determined from the extrapolated parts of the rating curve (Fig. 43).

The mean monthly water discharge is normally highest in July and lowest in March or April (Fig. 45). The run-off during the three summer months June—August normally amounts to more than 70 % of the total annual run-off. The run-off during the winter half-year is very low and normally amounts to only about 10 % of the total annual run-off.

The variation in water discharge during the summer half-year is illustrated in Fig. 46 for the years 1916, 1936 and 1954 by means of mean daily flows. In 1916 the water discharge was lower than normal during practically the whole summer, in 1936 it was higher than normal during the greater part of the summer, and in 1954 it was higher than normal during the greater part of May and August. As shown by the diagrams the variations in water discharge may be very great and very rapid, especially in summer but often also in spring and in autumn. During dry periods in summer with a marked daily variation in air temperature, there is also a very distinct daily variation in water discharge (Fig. 32). In winter, and especially in late winter, the variations in water discharge are normally very insignificant, since rain and melt water, produced during warmer periods in winter, mainly will be absorbed by the snow cover.

The average, the highest, and the lowest daily mean water discharge during the summer half-year is illustrated in Fig. 47 on the basis of oil-stage records at Litnok during the 30 years' period 1916—1945. This 30 years' period is bet-

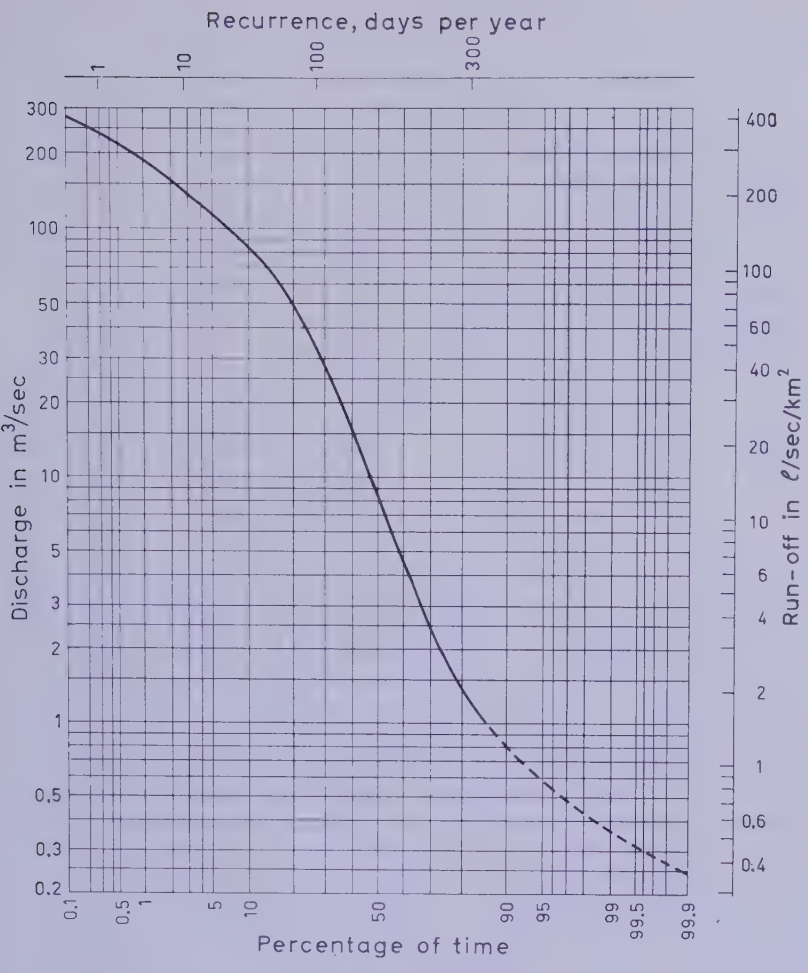


Fig. 48. Flow-duration curve for Rapaälven at Litnok, determined from mean daily flows (1916—1945).

ter covered by oil-stage records than the standard period 1931—1960. The oil-stage records cover 85 % of the period 1916—1945, which means in reality that the diagram is based on 23 to 29 daily means per date. As shown by the diagram the water discharge may be very high or rather low on every day in summer. Water discharges higher than the mean annual extreme value occurred during this period in June, July and August. The water discharge normally begins to increase more markedly in May when the snow begins to melt. The production of melt-water from the snow cover and from the glaciers, however, is often interrupted by cold weather. Periods with heavy rain are normally very short. The steep topography, the small thickness of the soil cover, and the lack of larger lakes favour a rapid run-off. The frequency of individual floods is therefore very high from May to September in relation to the size of the drainage basin, while the duration of the individual floods is very short.

The flow-duration curve (Fig. 48) is also based on mean daily flows during the period 1916—1945. Flows higher than the mean flow (27.4 m^3/sec.) have a total duration of about 31 % or about 112 days a year on average. The median flow as regards the duration (8.4 m^3/sec.) is about 3 times lower than the mean flow. The median flow as regards the total flow volume (79 m^3/sec., see Fig. 67) is about 3 times higher than the mean flow, and flows higher than this median flow have a total duration of about 11 % or about 40 days a year on average (see Fig. 66). Flows higher than the bankfull discharge in the proximal delta sec-

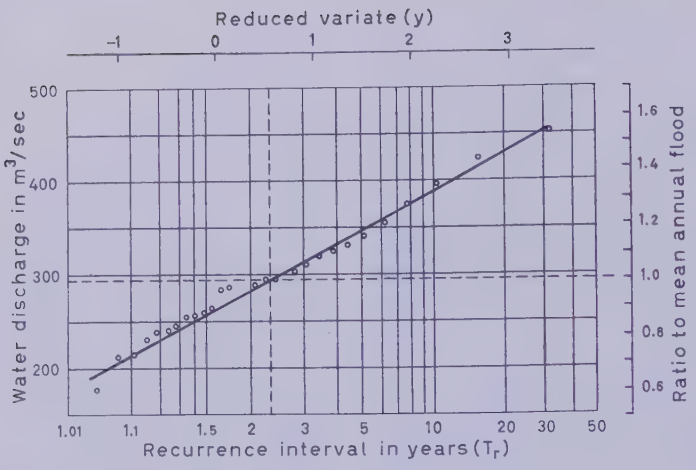


Fig. 49. Flood-frequency curve for Rapaälven at Litnok, determined from annual floods (1916—1945).

tion (about 280 m³/sec., see Fig. 67) have a total duration of only about 0.1 % (Fig. 48), and contribute only about 1 % of the total flow volume (Fig. 66). This duration of flows higher than the bankfull is considerably lower than the duration (0.6 %) that *Nixon* (1959) found to be the average for rivers in England and Wales. Flows higher and lower than the highest and the lowest directly measured flows (253 m³/sec. respectively 1.7 m³/sec.) have a total duration of about 0.2 % and about 24 % respectively, and contribute in total to only about 2 % of the total flow volume. This means that flows calculated on the basis of the less accurate, extrapolated parts of the rating curve (Fig. 43) have a very slight influence on the calculations of the total flow volume.

The flood-frequency curve (Fig. 49) is based on only the maximum flood peak of each calender year 1916—1945. For three of these years the annual peak occurred when the oil-stage recorder was out of operation. For these years the probable annual peak discharge has been estimated by comparison with the annual peak discharge in the river Blackälven at Tjåmotis 1. Recurrence intervals were computed by the formula

$$T_r = \frac{n+1}{m} \tag{19}$$

where T_r is the return period, or recurrence interval, in years, n the number of years of record, and m the rank starting with the highest as 1. This simple formula is applicable both to the annual-flood series and the partial-duration series, and gives results that acceptably conform with modern theories (*Dalrymple*, 1960; *Langbein*, 1960).

Hydrological extreme values have been found to be well approximated by a Fisher-Tippett type I distribution of extreme values, sometimes called a Gumbel distribution (*Fisher and Tippett*, 1928; *Gumbel*, 1954). According to *Chow* (1951) most frequency functions applicable for hydrologic analysis can be resolved to the generalized form

$$x = \bar{x} + K\sigma_x \tag{20}$$

where x is a variable, $\bar{x}$ the mean of x, and K a frequency factor. For the Gumbel distribution the frequency factor (a function of T_r and n) may be written

$$K = \frac{y - \bar{y}_n}{\sigma_n} \tag{21}$$

where y is a linear distance, termed the reduced variate, and $\bar{y}_n$ the mean of y for the particular sample size (n). The equation for the reduced variate is

$$y = -ln\left[-ln\left(1 - \frac{1}{T_r}\right)\right] \tag{22}$$

A plot of x (equation 20) versus y (equation 22) is linear on a Cartesian grid and forms the basis of the extremal probability paper introduced by *Powell* (1943). The nature of the Gumbel distri-

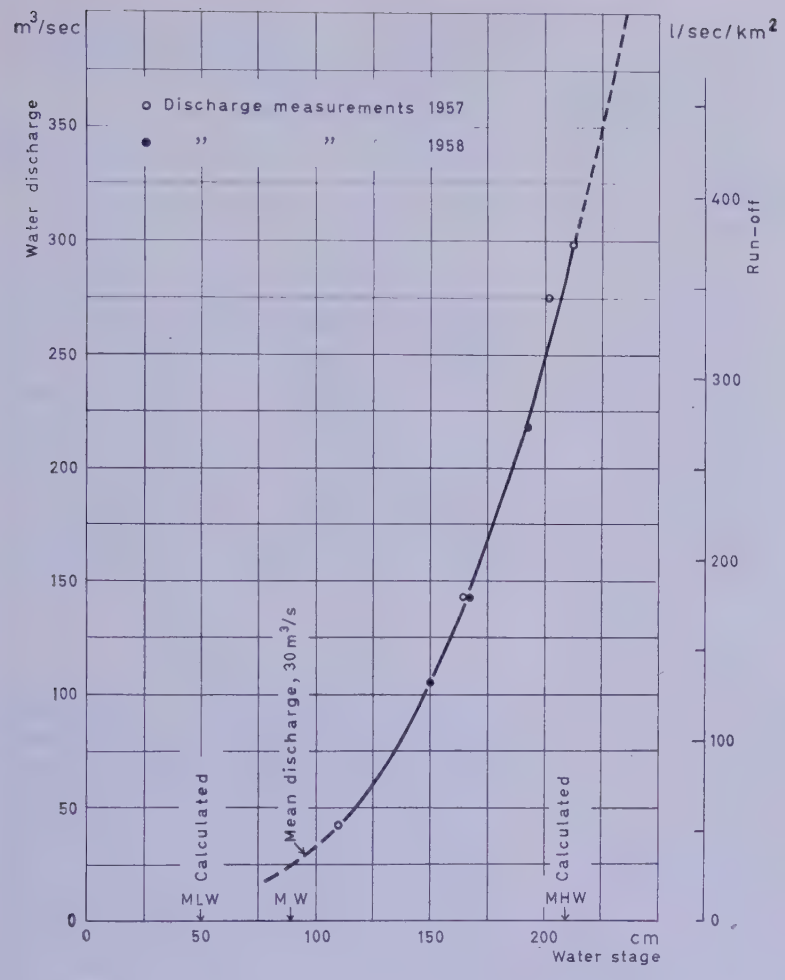


Fig. 50. Rating curve for Lake Laitaure.

bution is such that the arithmetic average of the annual series has a return period which approaches 2.33 years for large values of n (cf. *Benson*, 1960). Values of y corresponding to T_r are tabulated by, among others, *Gumbel* (1942). Nomograms for determining various return periods directly from the mean and standard deviation of the annual-maximum series are published by *Weiss* (1955, 1957).

The time graduation of the flood-frequency curve (Fig. 49) is based on equation (22). The mean value of the annual-flood series is 294 m³/sec. The standard deviation is relatively small and amounts to only 62 m³/sec. The 30-yr flood, for example, is therefore only 1.5—1.6 times higher than the mean annual flood. The bankfull discharge in the proximal delta section (about 280 m³/sec.) has a recurrence interval of 1.5—2 years according to the annual-flood series (about 1.4 years according to the partial-duration series). This recurrence interval of the bankfull discharge is about the same as *Wolman and Leopold* (1957) found for flood plains in general.

As shown by the diagram (Fig. 49) the recurrence intervals computed by formula (19) conform rather well with the Gumbel distribution. According to *Potter* (1958) annual-flood series plotted on extremal probability paper many times exhibit a dog leg, indicating that the samples of maximum annual peaks are drawn from two different populations, for example, peaks caused by rain and peaks caused by melting snow. The annual-flood series for Rapaälven does not exhibit any significant dog leg. It must, however, be

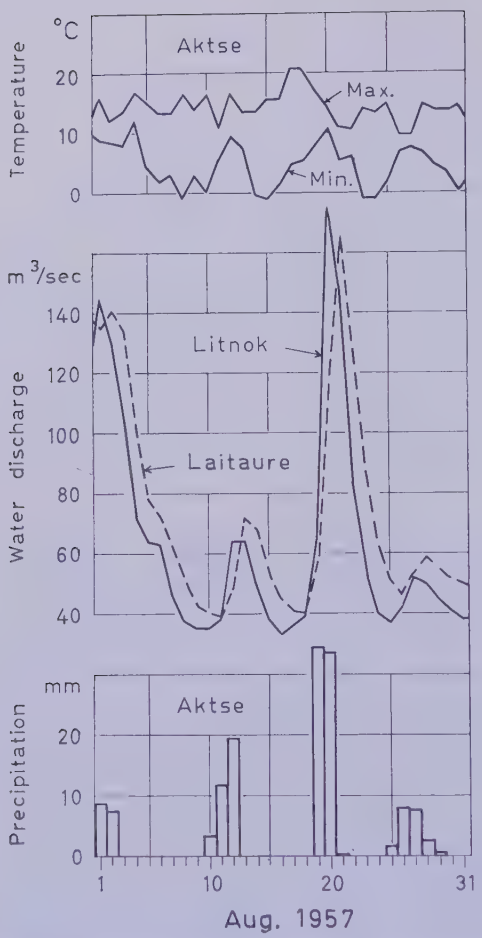


Fig. 51. Daily mean water discharge at Litnok and at the outlet of Lake Laitaure during August 1957. Temperature and precipitation at Aktse.

stressed that the higher floods are determined from the extrapolated part of the rating curve, and that the upper part of the flood-frequency curve must therefore be regarded as very approximate.

Laitaure

The rating curve for Laitaure (Fig. 50) is based on discharge measurements carried out in 1957 and in 1958 at the outlet section c—d (for location, see Plate II). The primary aim of these discharge measurements was to get the stage-discharge relation for those periods when the sediment load is high enough to be of greater importance for the calculations of the total annual sediment load at the outlet of the lake. The discharge measurements carried out by the author were therefore concentrated to high-water periods.

Discharge measurements during both high-water and low-water periods were later carried out at the outlet of Laitaure by officials of the Swedish Meteorological and Hydrological Institute.

The drainage area is about 19 % greater at the mentioned section than at Litnok. The mean annual water discharge ought to be about 30 m^3/sec. at this section, or about 10 % greater than at Litnok. Since normally the local inflow is relatively small compared with the main-stream inflow the variations in water discharge at the outlet of Laitaure mainly reflect the discharge variations in Rapaälven upstream of the lake (Fig. 51). The variations, especially the daily variations, however, are smaller and less rapid downstream than upstream of Laitaure due to the effect of storage.

Water temperature

The temperature of the water was generally read by means of thermistors and reversing thermometers. During shorter periods the daily variation in water temperature was recorded by bimetallic thermographs. Most of the temperature measurements were carried out in July and August, 1953, 1954, 1957 and 1958.

Since Rapaälven is fed with melt water throughout the summer the temperature of the river water is low also in this season and lower than the air temperature during the greater part of the summer. During the greater part of July and August the daily mean temperature of the river water at Litnok was higher than 6° but lower than 8°. For the period July—August, 1957 the mean temperature amounted to about 7° in Rapaälven at Litnok, to about 7.5° in the distributaries at the delta front, and to about 10° in the outlet from Laitaure.

During cloudless days in summer the temperature of the river water shows rather regular fluctuations from late afternoon maxima to early morning minima. In Rapaälven at Litnok the deviation from the daily mean temperature was usually greatest at about 6 p.m. and 7 a.m., and smallest at noon and midnight. This was, for example, the case during the period July 2—14, 1957. For this period the mean water temperature (6.5°) was 3.3° lower than the mean air temperature. The mean diurnal range in water tempera-

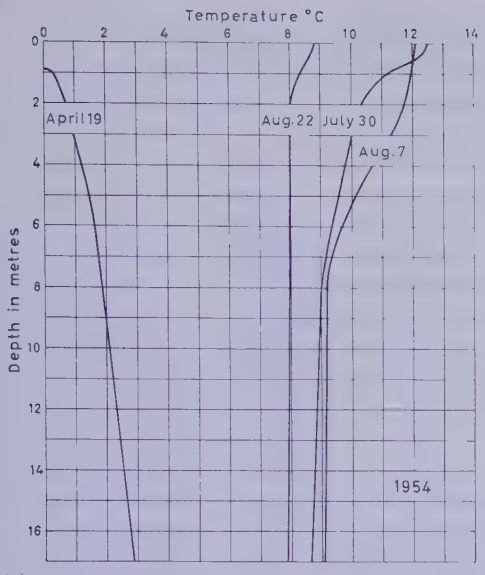


Fig. 52. Temperature profiles in the central, deepest part of Lake Laitaure in 1954.

ture (3.2°) was about 5 times smaller than the mean diurnal range in air temperature. For the period illustrated in Fig. 32 (August 3—10, 1953) the mean water temperature (7.0°) was 2.6° lower than the mean air temperature, while the mean diurnal range in water temperature (2.4°) was about 4 times smaller than the mean diurnal range in air temperature.

In those delta lakes that are cut off from channel connection with the distributaries the temperature of the surface water may be rather high during warmer periods in summer. In the shallower of these lakes the surface water may sometimes be more than 10° warmer than the river water in the adjoining distributaries. For the period July 18—24, 1959 the mean temperature of the surface water in the delta lake Saivajaurats amounted to 13.4°, while the mean temperature of the river water in the adjoining distributary Nuortätno amounted to 7.3° (for location, see Plate I).

In the central part of Laitaure the surface water ni summer is often 3° to 4° warmer than the inflowing river water, while the bottom water often is about 1° warmer than the unmixed river water. The thermocline is usually shallow and poorly developed. The temperature conditions in this part of the lake are exemplified in Fig. 52 by means of a few temperature profiles.

Dissolved solids

Theoretically, dissolved solids are the anhydrous residues of the dissolved substances in water. In reality, however, the term "dissolved solids" is usually defined as the solid residue remaining after evaporation of a filtered quantity of water to dryness (cf. *Hutchinson*, 1957; *Hem*, 1959; *Rainwater and Thatcher*, 1960; *Reid*, 1961). The difference between the residue on evaporation and the residue on ignition is called the loss on ignition. After correction for the CO_2 that is lost in evaporation and ignition the loss on ignition may be regarded as a rough estimate of the dissolved organic material (*Rodhe*, 1951).

The salinity of fresh waters may be regarded as the total concentration of the ionic constituents, and may usually be satisfactorily defined as the total concentration of the major ionic constituents: Ca, Mg, Na, K, CO_3, SO_4, and Cl (all bicarbonate being converted to carbonate). SiO_2, Al_2O_3 and Fe_2O_3 are often included in the definition of salinity, but according to *Hutchinson* (1957) most of the silicate is probably present as undissociated silicic acid and the iron and aluminium oxides are rarely in an ionic condition.

The specific conductance of water is easily and conveniently determined and gives an indication of the total ionic concentration. According to *Rodhe* (1949) ions other than the seven major constituents have, as a rule, very little influence on the conductance of lake waters. The relationship of salinity to conductance varies with the ionic composition and with the concentration. For bicarbonate waters of standard composition the relationship of salinity (as defined above) to specific conductance (in micromhos at 20°C) is about 0.52 at a salinity of 10 mg/1, about 0.58 at a salinity of 100 mg/1, and about 0.61 at a salinity of 200 mg/1 (see further *Rodhe*, 1949).

The content of dissolved solids (defined as the residue on evaporation at 105°C) is very low in Rapaälven and in Laitaure. In Rapaälven at the proximal delta section the discharge-weighted mean content (based on 22 analyses) amounts to 15 mg/1. This value corresponds to a dissolved-solids discharge of about 13,000 tons per year (about 1/12 of the suspended-sediment discharge), and to an average gross yield of about 19 tons per km^2 and year. Probably about one fifth of this gross yield consists of organic material.

The ionic composition of some of the analysed water samples is presented in Table 2. The percentage equivalent proportions of cations and

		Rapaälven, proximal delta section		Delta lake, 100 m WNW of Saivajaurats	Lake Laitaure, central part	Lake Laitaure, outlet section	
Date of collection		July 23	Aug. 6	Aug. 6	Aug. 6	Aug. 6	July 23
Water discharge, m^3/sec.		134	45			52	146
Water temperature, °C		7.2	7.5	15.0	11.9	10.3	10.1
Specific conductance, micromhos at 20°C		17.2	20.1	44.0	20.3	20.2	18.4
pH		6.8	7.1	6.6	7.0	7.1	7.0
Ca	mg/l	1.9	2.0	3.4	1.9	2.0	1.9
	mval-%	65	63	43	62	67	64
Mg	mg/l	0.2	0.1	0.9	0.1	0.1	0.2
	mval-%	11	5	19	5	6	11
Na	mg/l	0.5	0.8	1.4	0.8	0.6	0.5
	mval-%	15	22	15	23	17	15
K	mg/l	0.5	0.6	3.5	0.6	0.6	0.6
	mval-%	9	10	23	10	10	10
Cl	mg/l	0.2	0.3	0.3	0.4	0.2	0.2
	mval-%	4	5	2	7	3	4
SO_4	mg/l	2.8	2.9	0.3	3.1	3.4	2.8
	mval-%	37	35	1	39	38	35
HCO_3	mg/l	5.6	6.4	26.0	5.4	6.8	6.2
	mval-%	59	60	97	54	59	61
Residue on evaporation at 105°C, mg/l		13	16	35	15	17	15
Residue on ignition at 800°C, mg/l		8	11	13	10	11	9
Dissolved organic material, mg/l		3	3	12	3	4	4
Salinity, mg/l		8.9	9.8	22.6	9.6	10.2	9.2

Table 2. Surface-water samples collected and analysed in 1959.

anions were found to be rather constant in Rapaälven, but varied considerably in the small, shallow, and enclosed delta lakes. In those years when the levees are not topped by flood-water the salinity in these lakes may be more than twice as high as in the distributaries. For the river water the relationship of salinity to specific conductance was found to be 0.50 (mean of 14 analyses).

The specific conductance of the river water (determined with a portable field equipment) generally decreased with increasing water discharge and varied between 14 and 27 micromhos at 20°C. The discharge-weighted mean value, 19 micromhos at 20°C, corresponds to about 0.16 mval/l of cations and anions and to a salinity of 9—10 mg/l. This value of the salinity corresponds to a salt discharge of about 8,000 tons per year, and to an average gross yield of about 12 tons per km^2 and year upstream of the Laitaure delta. The major part of this gross yield is probably derived from the atmosphere through wet and dry deposition.

Chapter 9

TRANSPORTATION OF SEDIMENT

Bed load

Many attempts have been made to determine the bed-load discharge in rivers by direct measurement, with some type of apparatus, and by the use of empirical or theoretically deduced formulas. None of the proposed methods, however, has been universally accepted as completely adequate for the determination of bed-load discharge in natural river channels (*Colby*, 1963; *Hubbell*, 1964).

For computing the bed-load discharge in Rapaälven, at the proximal delta section, the bed-load equation developed by *Kalinske* (1947) was found to be suitable. Computations by this equation usually give bed-load discharges in the proper order of magnitude under a wide variety of field and laboratory conditions (*Sundborg*, 1956; *Hubbell and Matejka*, 1959; *Inter-Agency Committee on Water Resources*, 1963). *Kalinske's* bed-load equation may be written

$$l_b = 7.3\gamma d\varrho_s \sqrt{\frac{\tau_0}{\varrho_f}}\, F\left(\frac{\tau_c}{\tau_0}\right) \qquad (23)$$

where l_b is the rate of bed-load transport per unit width of the stream bed, expressed as a weight per unit time ($g \cdot cm^{-1} \cdot sec^{-1}$); γ a factor which indicates the proportion of the bed taking the fluid shear (about 0.35 for unigranular material according to data obtained by *White*, 1940); ϱ_s the density of the sediment particles; τ_c the critical value of the boundary shear, i.e. the value at which bed-load transport commences; and $F(\tau_c/\tau_0)$ a special function for bed-load transport, graphically illustrated in Fig. 2 of *Kalinske's* paper. The function is dependent on the state of turbulence.

The results of computations based on *Kalinske's* equation and necessary field data, collected on 6 occasions, are plotted in Fig. 53. Values of the critical boundary shear were computed by formulas developed by *Sundborg* (1956). The relationship between the water discharge and the bed-load discharge in Rapaälven, at the proximal delta section, was found to be well fitted by a straight line. The equation for the rating curve is

$$L_b = 4.63 \cdot 10^{-2}(Q_R - 20) \qquad (24)$$

where L_b is the bed-load discharge in kg/sec. in Rapaälven, at the proximal delta section.

According to the rating curve transportation of bed load commences in the studied section when the water discharge is 20 m^3/sec. This value is in fair agreement with the direct observations. General transportation of bed load was observed to take place when the water discharge was higher than 26 m^3/sec. No transportation of bed load was observed when the water discharge, during the ice-free period, was lower than 18 m^3/sec. In winter transportation of bed load may take place at rather low water discharges if the effective cross-sectional area is reduced due to the formation of frazil and anchor ice. The effect of ice-damming, however, is probably very slight in this part of the river. Flows higher than 20 m^3/sec. have on an average a total duration of about 36 % or about 130 days a year. General transportation of bed load in this section ought therefore to take place during, on an average, only about one third of a year.

Calculations based on equation (24) and water-discharge data for the 30 years' period 1916—1945 give the result that the bed-load discharge in the proximal delta section amounts to, on an average, about 25,000 metric tons a year. This value corresponds to about 16 % of the suspended-load discharge and to about 14 % of the total sediment discharge in this section. The computed relationship between the average bed-load discharge and the average suspended-load discharge seems to be in the proper order of magnitude.

Suspended sediment

The concentration and discharge of suspended sediment was studied in more detail in the proximal delta section and in the outlet section of Laitaure. Most of the water samples were taken in vertical I of these sections with point-integrating samplers (see further *Axelsson*, 1956).

The concentration of suspended sediment was

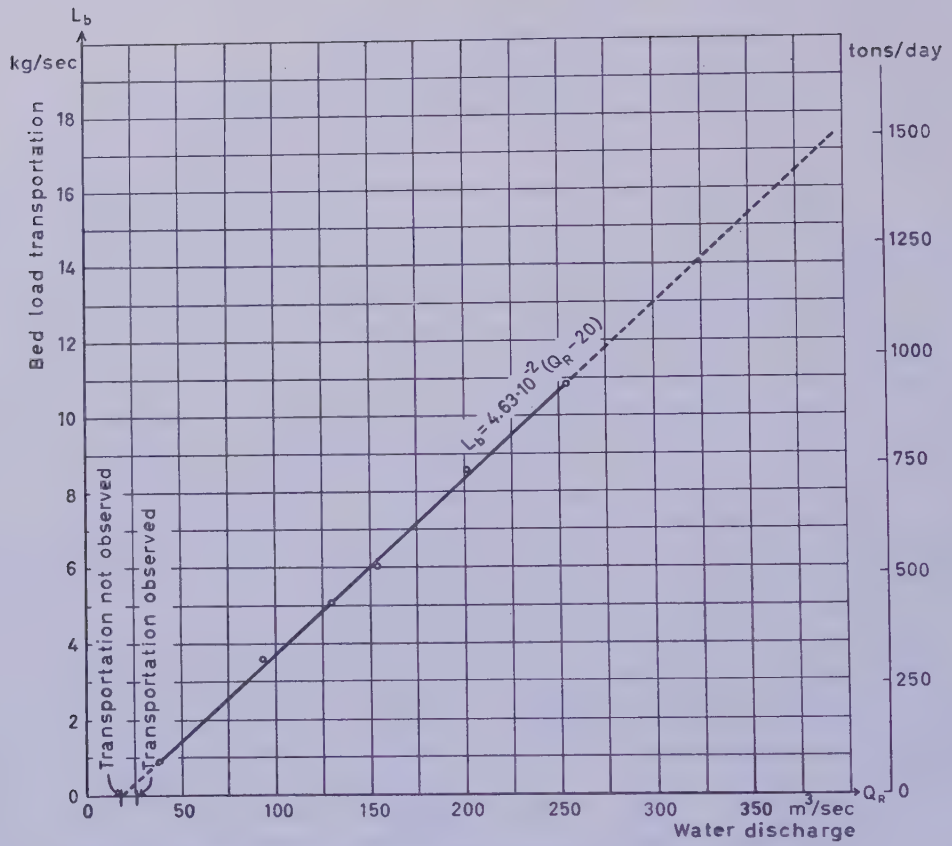


Fig. 53. Bed-load discharge rating curve for Rapaälven at the proximal delta section.

determined by the filter-paper method. The water samples were filtered through Munktell's filter paper No. 00 with a diameter of 18.5 cm and an ash weight of 0.12 mg. The filter paper was then ignited at a temperature of 800° C, and the residue weighed by means of an analysis balance with an accuracy of 0.1 mg. The ordinary filter-paper method is not suitable when the suspended sediment is very fine-grained. In Rapaälven and also in Laitaure, however, the suspended sediment is relatively coarse-grained, especially when the sediment concentration is high (cf. Figs. 55 and 61). It was therefore considered justifiable to use filter paper of the type mentioned. It was also found that the amount of suspended sediment that passed through the filter paper was negligible.

Most of the water samples taken in the summer of 1957 were collected by *L. Lund* (at the proximal delta section) and *S. Zackrisson* (at the outlet of Laitaure), who assisted the author during the field work this summer, and who have also reported some preliminary results in unpublished examination papers (*Lund*, 1959; *Zackrisson*, 1957).

The concentration and discharge of suspended sediment in Rapaälven was earlier studied by *Hamberg*. His calculations of the suspended-sediment discharge were based on water samples taken on 4 occasions in the 1890's, just downstream of the proximal delta section (*Hamberg*, 1901).

Sediment concentration

According to *Rouse* (1937) the vertical distribution of suspended sediment may be presented by the formula

$$\frac{c_y}{c_a} = \left(\frac{D-y}{y} \cdot \frac{a}{D-a} \right)^{z_t} \qquad (25)$$

where c_y is the concentration of sediment of a given grain size at a distance y above the bed, c_a the concentration of sediment of the same grain size at a distance a above the bed, D the water depth, and z_t the theoretical exponent of the vertical distribution of suspended sediment of a given grain size. The theoretical exponent of suspended-sediment distribution may be written

$$z_t = \frac{v_s}{\beta k u_*} \tag{26}$$

where v_s is the settling velocity of the sediment particles, and β the ratio of the transfer coefficient for sediment to the momentum transfer coefficient (usually close to unity).

The formula (25) has been deduced on the assumption that a state of equilibrium prevails as regards the concentration of suspended sediment. The formula is not valid if the concentration of sediment is so high that density stratification occurs, and does not apply at the water surface, where the computed concentration would be zero, or at the bed, where the computed concentration is infinite. Measurements in natural streams have shown that the exponential form of the formula (25) is satisfactory but that the suspended sediment usually is distributed more uniformly than conventional theory predicts. Whenever possible z_t therefore should be replaced by z_1, the exponent that fits the measured *data*, equal to the slope of the log-log plot of $(D-y)/y$ versus the concentration of a given grain size or grain-size fraction (*Vanoni*, 1941; *Andersson*, 1942).

The deviation of z_1 from z_t is usually small for relatively fine-grained particles but becomes greater with increasing particle size (*Einstein and Chien*, 1954; *Chien*, 1956). *Hubbell and Matejka* (1959) found that z_1 varies inversely with about the 0.6 power of k for streams having a range of k from 0.4 to 4.0. *Colby and Hembree* (1955) found that z_1 varies, on an average, with about the 0.7 power of the fall velocity if the fall velocities computed by *Rubey* (1933) were used. *Nordin and Dempster* (1963), using the fall velocities from Fig. 5 of Report No. 4 from *Inter-Agency Committee on Water Resources* (1941), found that z_1 varies with about the 0.55 power of the fall velocity. In addition z_1 is sometimes expressed as a power of z_t; the power, however, depends partly on which system of fall velocities is used (*Hubbell and Matejka*, 1959).

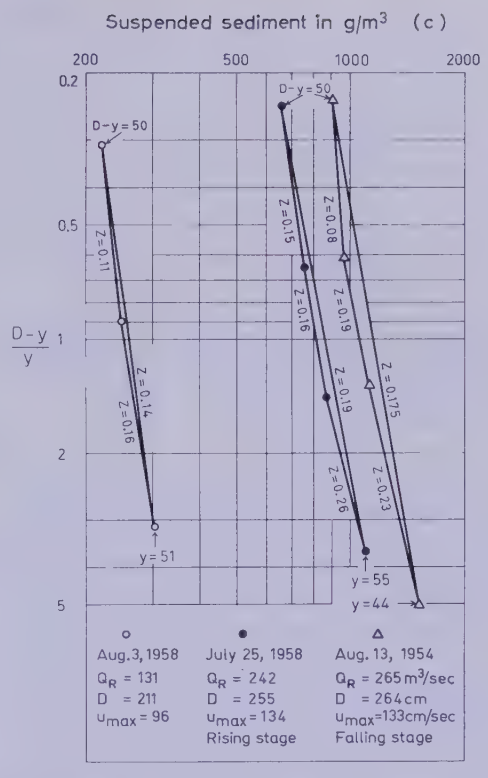


Fig. 54. Suspended-sediment distribution in vertical I of the proximal delta section.

Rapaälven at the proximal delta section

The concentration of various grain-size fractions in vertical I of the proximal delta section was in general too low to allow a correct determination of z_1 on the basis of the ordinary point-integrated samples, having a volume of only 1 litre. Instead of the concentration of various grain-size fractions the total concentration of suspended sediment therefore was plotted against $(D-y)/y$ as exemplified by Fig. 54. With z in Fig. 54 is thus meant the average slope of the log-log plot of $(D-y)/y$ versus the total concentration of occurring grain sizes at various depths in the vertical. It was found, as expected, that this z increased with depth, with increasing water discharge, and that it generally was higher during rising than during falling stages.

That also the median grain size of the suspended sediment increased with depth, with increasing water discharge, and that it generally was higher during rising than during falling

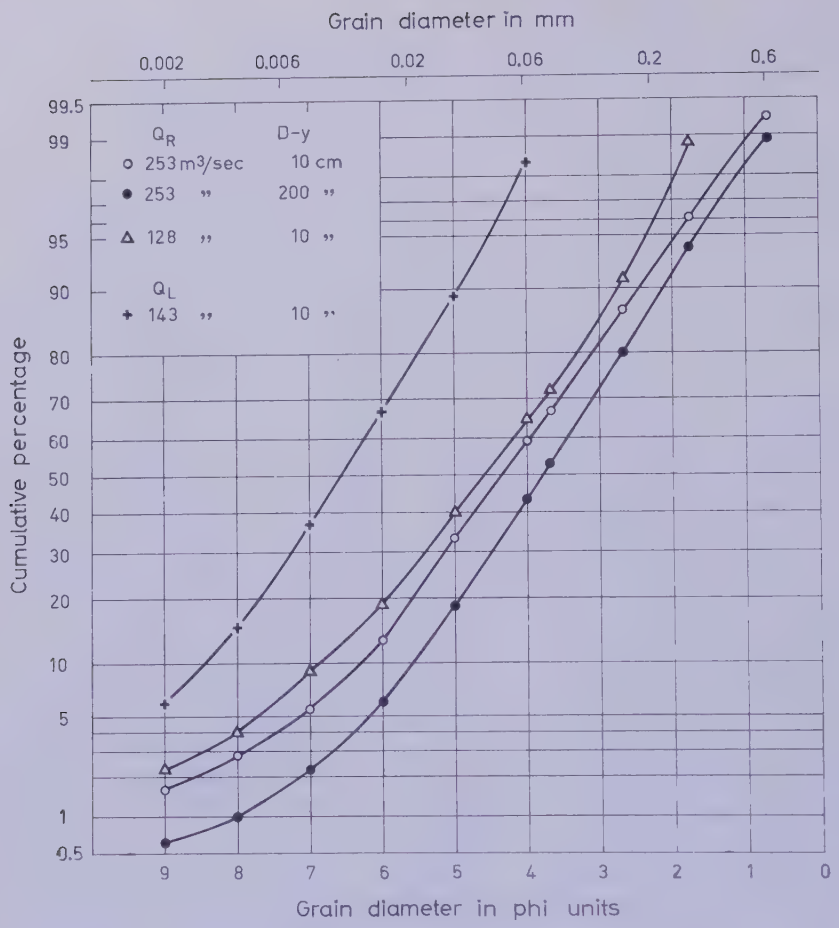


Fig. 55. Grain-size distribution of suspended sediment in vertical I of the proximal delta section (Q_R) and in vertical I of the outlet section of Lake Laitaure (Q_L). July, 1957.

stages was confirmed by the grain-size analyses. The grain-size distribution of some suspended-sediment samples is shown in Fig. 55. The grain-size analyses also gave the result that the median grain diameter of the total suspended-sediment load in the proximal delta section is about 0.05 mm, that about 8 % (by weight) is coarser than 0.2 mm, that about 80 % is coarser than 0.02 mm, and that only about 2 % is finer than 0.002 mm.

The grain-size distribution of suspended and deposited sediments was determined by sieving and by pipette analysis. Material that passed the 0.0615 mm sieve was taken for pipette analysis and was dispersed in distilled water with sodium oxalate as dispersing agent. As previously pointed out the salinity is very low in Rapaälven. The concentration of suspended material is also relatively low. It is therefore probable that the fine-grained suspended material is also rather well dispersed in the river water.

The relative sediment discharge at different depths in vertical I of the proximal delta section was computed by multiplying the flow velocity in percentage of the mean velocity in the vertical (u_p) by the suspended-sediment concentration (c) and by dividing the product with 100, as illustrated by Fig. 56. The discharge-weighted mean sediment concentration ($\bar{c}$) was then obtained by graphic integration. For those occasions when only samples of suspended sediment were taken at selected points in this vertical the probable vertical velocity distribution was determined graphically by means of the diagram, Fig. 41.

An increase in stream flow is usually accompanied by an increase in sediment concentration. The sediment-concentration peak asso-

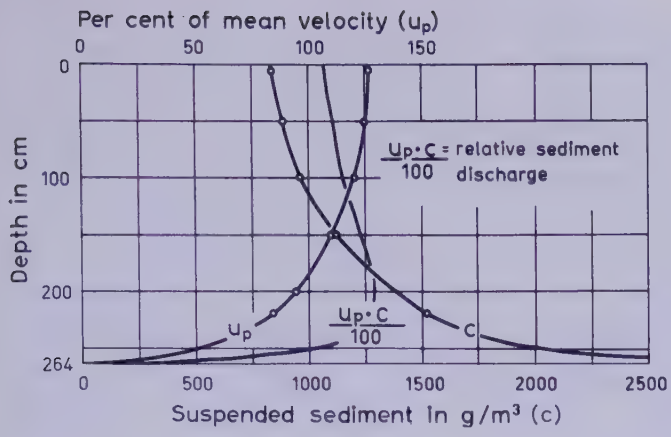


Fig. 56. Distribution of velocity, sediment, and sediment discharge in vertical I of the proximal delta section. August 13, 1954. Water discharge: 265 m³/sec. Falling stage.

ciated with a rise may precede, coincide with, or follow the water-discharge peak (see further *Heidel*, 1956). In Rapaälven the time-lag between the peak in sediment concentration and the peak in water discharge generally was very short. Sometimes the sediment-concentration peak preceded, sometimes it followed the water-discharge peak, but often the two peaks practically coincided. The sediment concentration was generally lower in spring, when the increase in stream flow was caused mainly by melting snow, than at corresponding water discharges in summer and in autumn; but the difference was not very pronounced.

The primary aim of the sediment-transport studies was to ascertain the average amount of

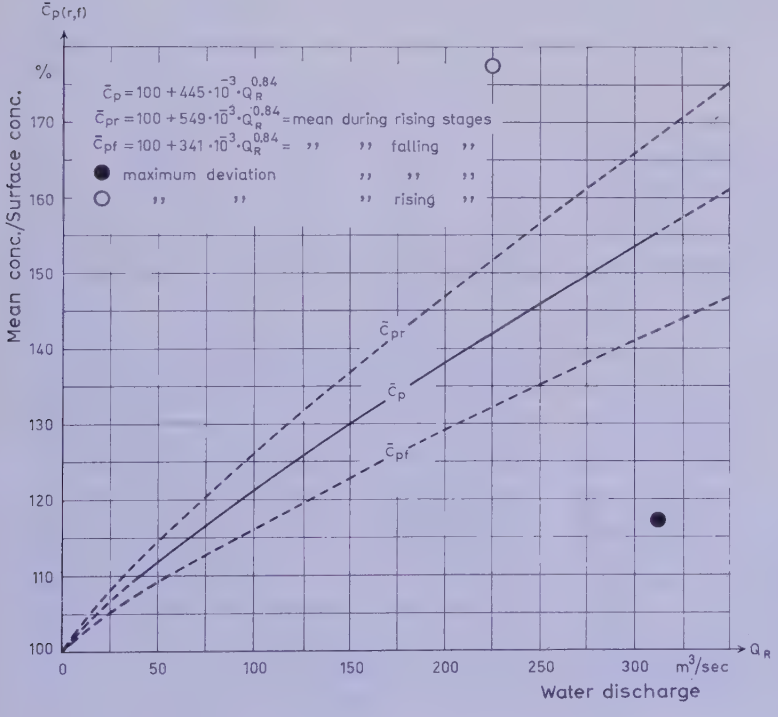


Fig. 57. Percentage relation of weighted mean sediment concentration to sediment concentration 5 cm below the water surface in vertical I of the proximal delta section, given as a function of the water discharge in the section.

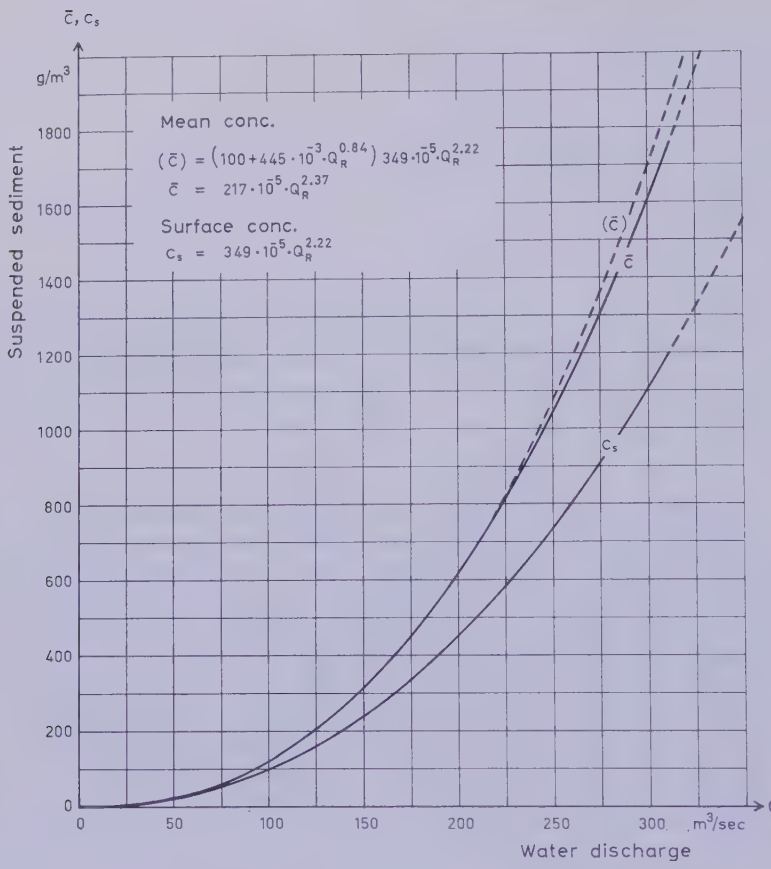


Fig. 58. The relationship between the sediment concentration in vertical I of the proximal delta section and the water discharge in the section.

sediment transported to the Laitaure delta during different water discharges. It was therefore found suitable to investigate the empirical relationship between sediment concentration and water discharge, and to illustrate the concentration and discharge of sediment as a function of the water discharge.

As illustrated by Fig. 57, the difference between the weighted mean sediment concentration, and the sediment concentration near the water surface in vertical I of the proximal delta section, generally increased with increasing water discharge and generally was greater during rising than during falling stages. The surface-water samples were generally taken 5 cm below the water surface. The average percentage relationship of weighted mean sediment concentration to sediment concentration near the water surface ($\bar{c}_p$) in the studied vertical (based on sampling on 41 occasions and calculated by the method of least squares) is given by the equations

$$\bar{c}_p = 100 + 445 \cdot 10^{-3} \cdot Q_R^{0.84} \quad (27)$$

$$\bar{c}_{pr} = 100 + 549 \cdot 10^{-3} \cdot Q_R^{0.84} \quad (28)$$

$$\bar{c}_{pf} = 100 + 341 \cdot 10^{-3} \cdot Q_R^{0.84} \quad (29)$$

where $\bar{c}_{pr}$ and $\bar{c}_{pf}$ is the weighted mean sediment concentration in percentage of the sediment concentration near the water surface, during rising and falling stages respectively. The deviations from the average relationships could sometimes be rather great as shown in Fig. 57.

For those occasions when only surface-water samples were taken in vertical I the weighted mean sediment concentration was estimated graphically by means of the diagram, Fig. 57.

The calculated relationship between the sediment concentration near the water surface in g/m³ (c_s) in vertical I and the water discharge in the section is based on surface-water samples taken on 286 occasions (1953—1960), and is given by the equation

$$c_s = 349 \cdot 10^{-5} \cdot Q_R^{2.22} \quad (30)$$

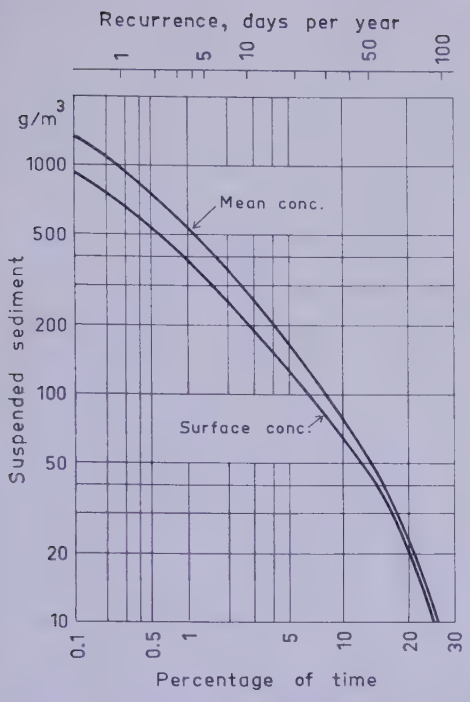


Fig. 59. Duration curves of suspended-sediment concentration in vertical I of the proximal delta section.

The relationship between the weighted mean sediment concentration in g/m³ ($\bar{c}$) in vertical I and the water discharge in the section is given by the equations

$$\bar{c} = 217 \cdot 10^{-5} \cdot Q_R^{2.37} \qquad (31)$$

$$(\bar{c}) = (100 + 445 \cdot 10^{-3} \cdot Q_R^{0.84})\, 349 \cdot 10^{-5} \cdot Q_R^{2.22} \qquad (32)$$

Equation (31) is based on equation (33), equation (32) on the equations (27) and (30). The equations (30), (31) and (32) are graphically illustrated in Fig. 58.

The suspended-sediment concentration may be very low in winter. The lowest measured sediment concentrations are lower than 1 mg/l. The highest measured mean concentration in vertical I amounts to 1260 mg/l. The highest measured concentration 5 cm below the water surface amounts to 1014 mg/l. According to equation (30) the sediment concentration near the water surface should be, on an average, about 950 mg/l at bankfull discharge in the section (about 280 m³/sec.). The measured sediment concentrations indicate that this value is in the proper order of magnitude. The measured concentrations at water discharges higher than the bankfull discharge are few, and performed mainly during the falling stage. They indicate, however, that the sediment concentrations given by the equations (30), (31) and (32) generally are too high for water discharges higher than the bankfull discharge.

The duration curves of suspended-sediment concentration (Fig. 59) are based on the equations (30) and (31) and on the flow-duration curve (Fig. 48). As shown by Fig. 59 the average duration of high sediment concentrations is short. Discharge-weighted concentrations higher than the discharge-weighted annual mean concentration (185 mg/l) have, on an average, a total duration of only about 4 % or about 15 days a year. During the greater part of the year the sediment concentration is lower than 10 mg/l.

The transverse distribution of suspended sediment in the proximal delta section was studied on 5 occasions by sampling in 7 verticals. It was found that the distribution in general was relatively uniform, as exemplified in Fig. 38 by weighted mean concentrations in the studied verticals. As shown by Fig. 60 the difference between the discharge-weighted mean concentration in the cross section and that in vertical I

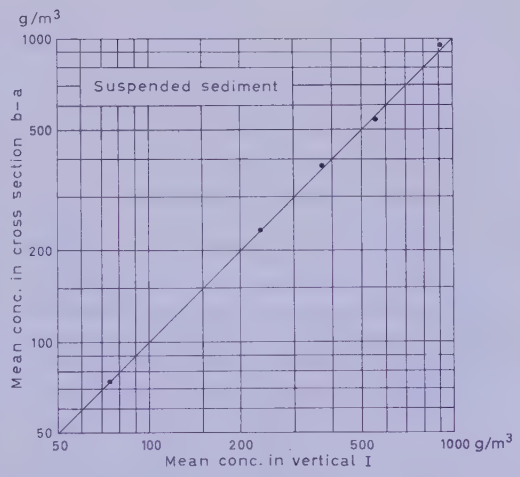


Fig. 60. The relationship between the weighted mean sediment concentration in vertical I of the proximal delta section and the weighted mean sediment concentration in the cross section.

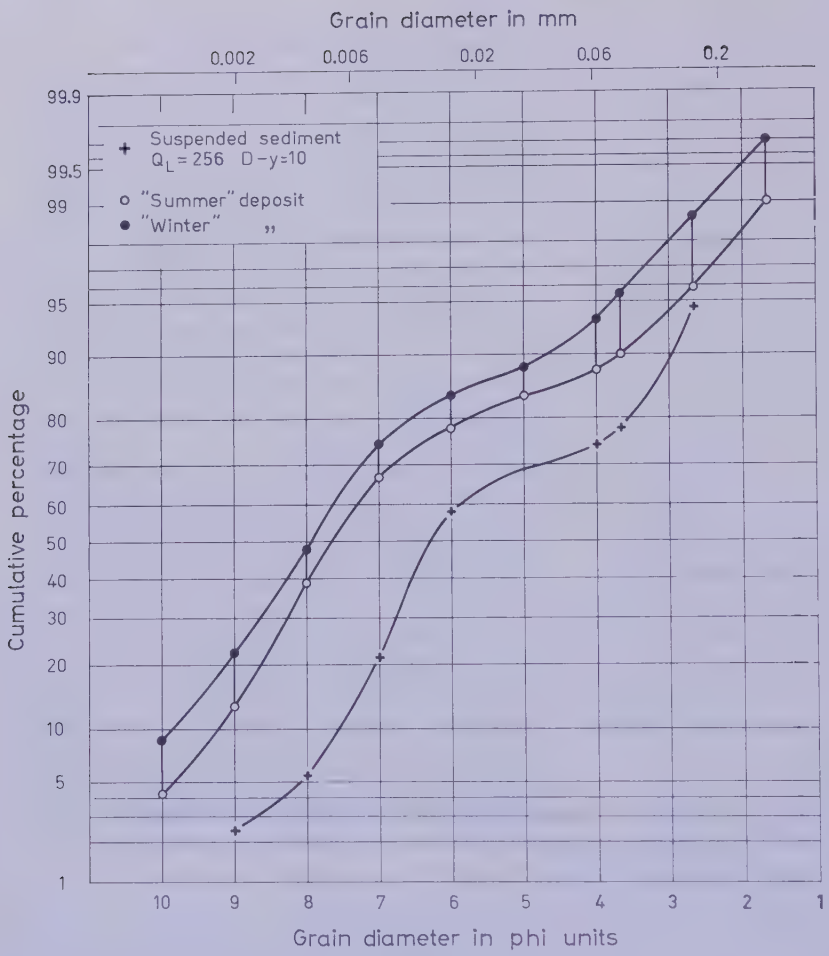


Fig. 61. Grain-size distribution of suspended and deposited sediment in vertical I of the outlet section of Lake Laitaure. July 20, 1957.

generally was insignificant. The weighted mean concentrations in vertical I are therefore taken as representative also of the cross section.

Laitaure at the outlet section

In the outlet of Laitaure the suspended sediment is considerably finer grained than upstream of the lake, as exemplified by Fig. 55. The vertical distribution of suspended sediment is therefore more uniform, and the deviation of the sediment concentration in the surface water from the weighted mean concentration is often small.

The upper parts of the varved deposits, in the deeper part of the outlet section, are characterized by a bimodal grain-size distribution (Fig. 61). The grain-size distribution of the suspended sediment is sometimes also bimodal, as shown in Fig. 61, probably due to erosion of coarser particles (probably sandy till) upstream of the section.

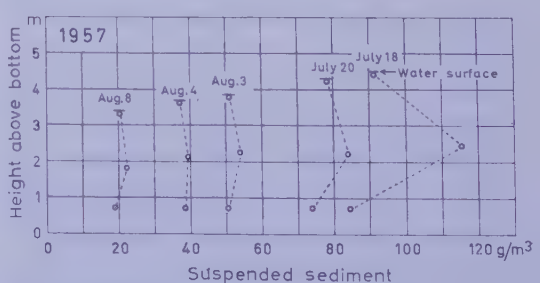


Fig. 62. Sediment concentration in vertical I of the oulet section of Lake Laitaure.

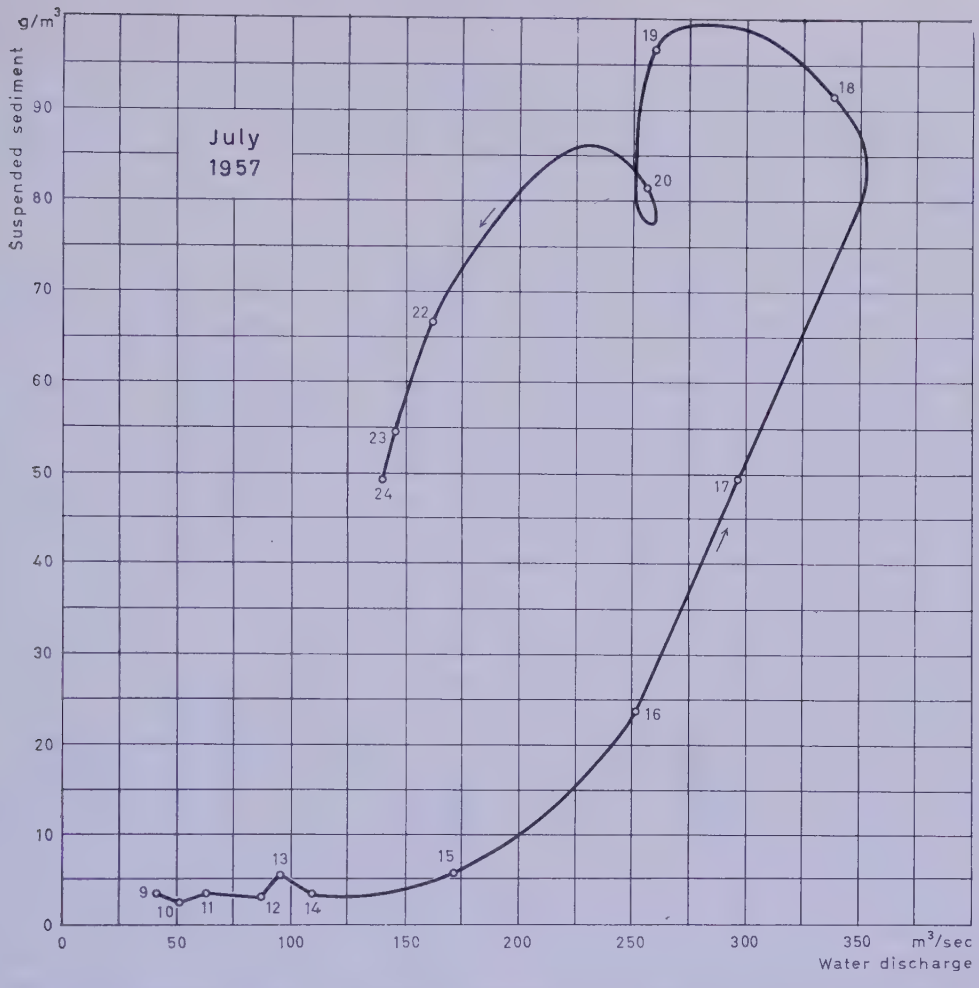


Fig. 63. The relationship between the sediment concentration, 5 cm below the water surface in vertical I of the outlet section of Lake Laitaure, and the water discharge in the section during the period July 9—24, 1957.

The bottom vegetation, in and upstream of the outlet section, not only affects the vertical distribution of velocity (Fig. 42) but also the vertical distribution of suspended sediment (Fig. 62). As shown in Fig. 62 the sediment concentration is sometimes lower in the bottom water than in the surface water.

Since the flood waves travel at a greater rate through the lake than the water and the suspended sediment, the peak in sediment concentration lags behind the peak in water discharge at the outlet of Laitaure. The sediment concentration may, therefore, be considerably lower during the rising than during the falling stage, as shown in Fig. 63.

In vertical I of the section the highest measured sediment concentration in the surface water amounts to 97 mg/l. The distribution of suspended sediment in the section is usually rather uniform. When computing the sediment discharge in the section the sediment concentration 5 cm below the water surface in vertical I is taken, therefore, as representative of the discharge-weighted mean concentration in the section.

Suspended-sediment discharge

Rapaälven at the proximal delta section

The discharge of suspended sediment in the proximal delta section was computed by multi-

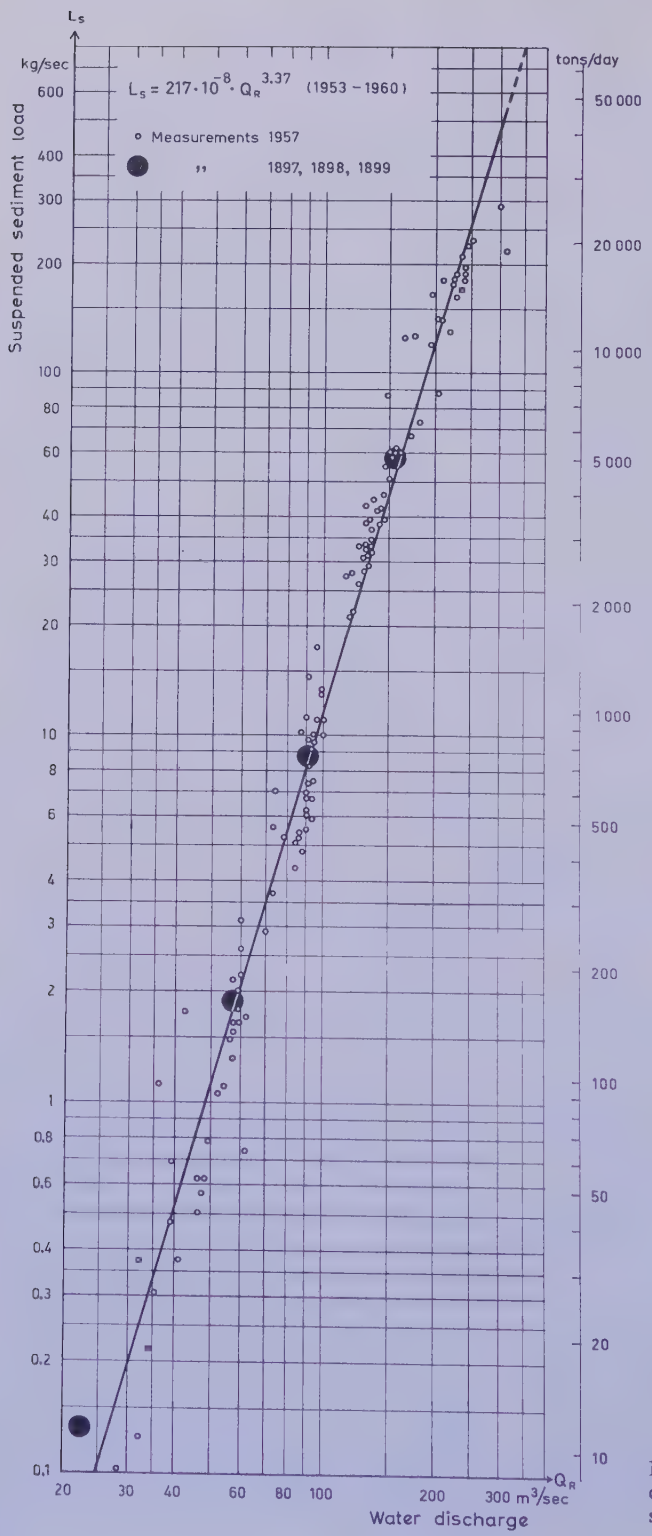


Fig. 64. Suspended-sediment discharge rating curve for Rapaälven at the proximal delta section.

plying the weighted mean sediment concentration by the water discharge. The relationship of suspended-sediment discharge to water discharge was then calculated by the method of least squares. For the period 1953—1960 the following relationship was obtained

$$L_s = 217 \cdot 10^{-8} \cdot Q_R^{3.37} \qquad (33)$$

where L_s is the suspended-sediment discharge in kg/sec.

Suspended-sediment discharges measured in 1957 are, in Fig. 64, plotted against water discharge. As shown by Fig. 64 the relationship of suspended-sediment discharge to water discharge is good, and very good in relation to most rivers. The rating curve, however, probably gives too high sediment discharges for water discharges higher than bankfull discharge (280 m^3/sec.).

The values of the suspended-sediment discharge in the 1890's, plotted in Fig. 64, are based on the values of water discharge and sediment concentration in the surface water given by *Hamberg* (1901). The values given by *Hamberg* are based on measurements on 4 occasions just downstream of the proximal delta section. The probable discharge-weighted mean concentrations on these occasions are estimated by means of equation (27). As shown by Fig. 64 the relationship of suspended-sediment discharge to water discharge was about the same in the 1890's as in the 1950's. It was therefore considered justifiable to use the rating curve, as defined by equation (33), for calculation of the probable discharge of suspended sediment even during periods not covered by the investigations, and to base the calculations on water discharge *data* for the 30-years' period 1916—1945.

According to the calculations the discharge of suspended sediment amounts, on an average, to 160,000 metric tons/year. This value is in close agreement with the average value (180,000 tons/year) roughly estimated by *Hamberg* for the period 1898—1900. Suspended-sediment discharges lower than 100,000 tons/year and higher than 200,000 tons/year probably occur 1 out of 5 years on an average.

The calculated bed-load discharge amounts, on an average, to about 25,000 tons/year. The average total sediment discharge thus ought to be about 185,000 tons/year. This value of the sediment discharge is very high in comparison with most rivers in Sweden.

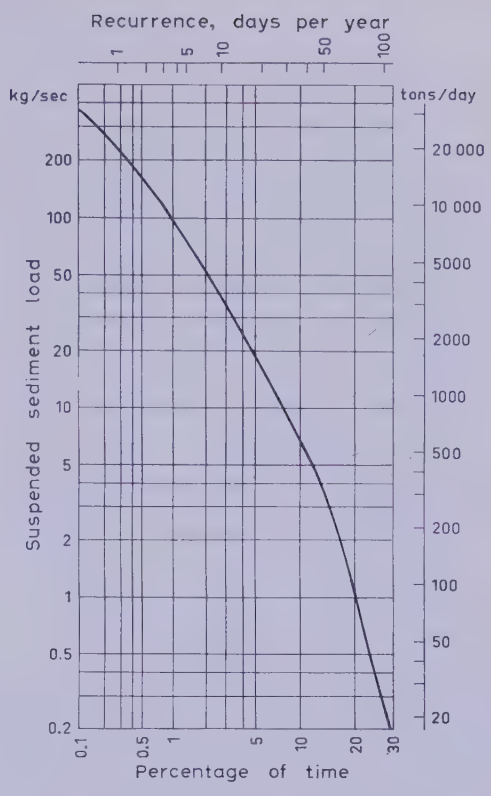


Fig. 65. Duration curve of suspended-sediment discharge in Rapaälven at the proximal delta section.

The annual variation in suspended-sediment discharge is extremely great. The mean annual extreme values are lower than 0.01 ton/day and higher than 20,000 tons/day respectively. The highest measured suspended-sediment discharge, 325 kg/sec., corresponds to a discharge of about 28,000 tons/day. About 95 % of the total annual suspended-sediment load is discharged during the three summer months June—August.

The duration curve of suspended-sediment discharge (Fig. 65) is based on equation (33) and on the flow-duration curve (Fig. 48). The average duration of high suspended-sediment discharges is short. Suspended-sediment discharges higher than the annual mean discharge, 5 kg/sec., have a total duration, on an average, of only about 12 % or about 44 days a year.

The cumulative percentage of water discharge, bed-load discharge, and suspended-sediment discharge are, in Fig. 66, plotted against cumulative percentage of time, and in Fig. 67 plotted against

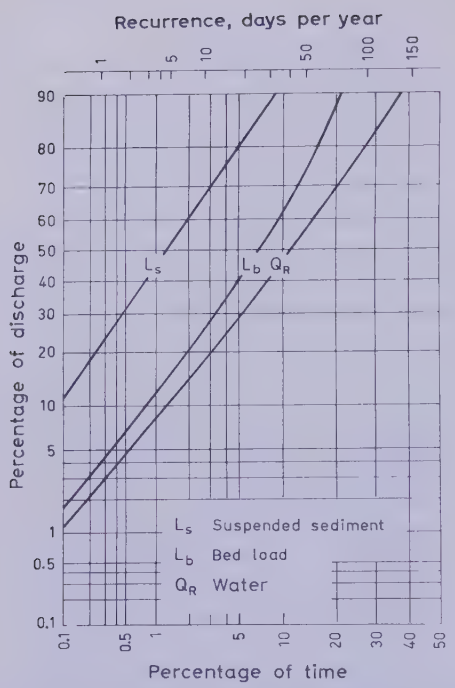


Fig. 66. Cumulative percentage of suspended-sediment discharge, bed-load discharge, and water discharge in the proximal delta section plotted against cumulative percentage of time.

water discharge. As shown by Fig. 66, half of the total annual discharge of suspended sediment takes place, on an average, during a total period of about 5 days. This means that some years half of the total annual delta growth may take place during a single week. As shown by Fig. 67, flows higher than 180 m^3/sec. contribute about half of the total annual discharge of suspended sediment. Flows lower than the mean discharge and higher than the bankfull discharge contribute in total only about 11 % of the total annual discharge of suspended sediment. This means that suspended-sediment discharges, calculated on the basis of the less accurate parts of the rating curve (Fig. 64), have a rather slight influence on the calculations of the total discharge of suspended sediment.

Laitaure at the outlet section

Because the peak in sediment concentration lags behind the peak in water discharge at the outlet of Laitaure, the suspended-sediment discharge shows a low correlation to water discharge. This is evident from Fig. 68, where suspended-sediment discharges, measured in 1957, are plotted against water discharge. For the period 1957—1960 the following relationship was obtained

$$L_s = 386 \cdot 10^{-7} \cdot Q_L^{2.3} \qquad (34)$$

According to calculations based on equation (34) the discharge of suspended sediment at the outlet section probably amounts, on an average, to about 17,000 tons/year. This means that about 90 % of the sediment discharged to the Laitaure delta is deposited upstream of the outlet section of Laitaure; most of it is probably deposited at the proximal part of the delta front.

The deposition of suspended sediment may be predicted by the sedimentation formula worked out by *Sundborg* (1956, 1958, 1964). This formula may be written

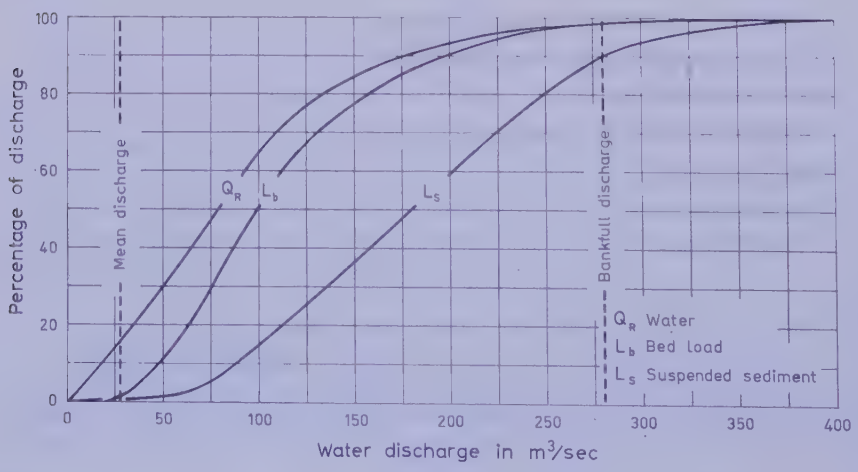


Fig. 67. Cumulative percentage of water-discharge, bed-load discharge, and suspended-sediment discharge in the proximal delta section plotted against water discharge.

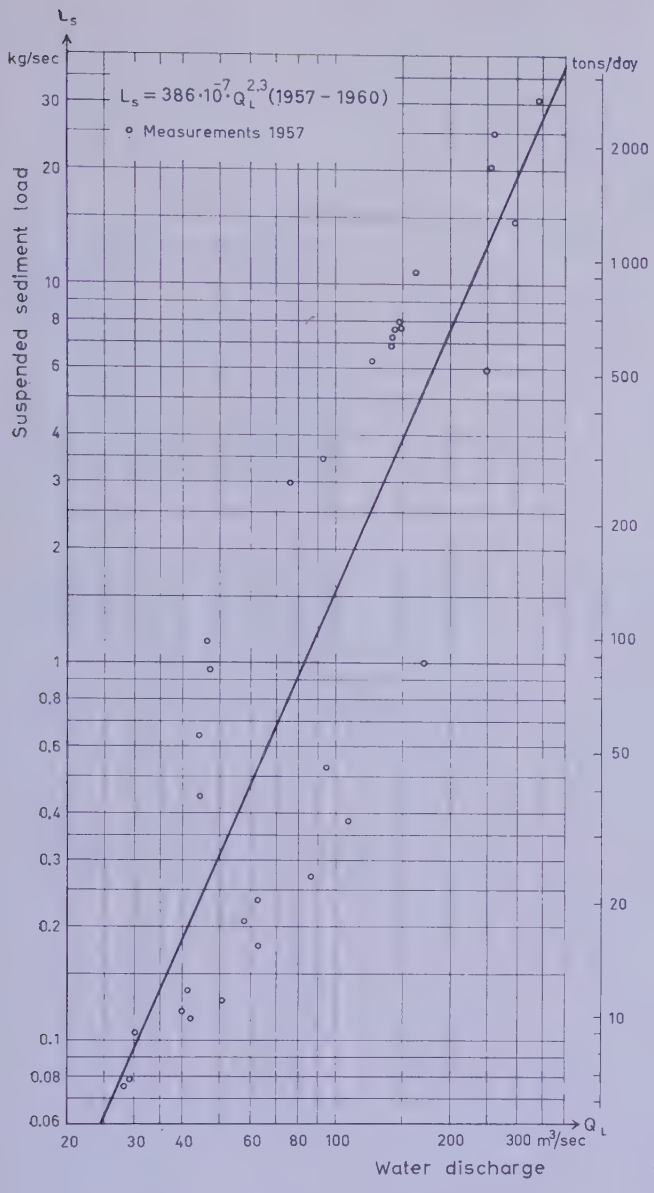


Fig. 68. Suspended-sediment discharge rating curve for Lake Laitaure at the outlet section.

$$\frac{\bar{c}_t}{\bar{c}_0} = e^{-v_s \Phi(v_s) \cdot t/D_m} \qquad (35)$$

where $\bar{c}_0$ and $\bar{c}_t$ is the mean vertical concentration of suspended sediment of a particular grain size at the time 0 and after the water has flowed for a period t respectively, e the base of Napierian logarithms, $\Phi(v_s) \cdot \bar{c}$ the concentration of the material in question immediately above the bed, and D_m the mean depth.

The deposition of suspended sediment in Laitaure, upstream of the outlet section, calculated by means of formula (35), is graphically illustrated in Fig. 69. The calculations are based on a water temperature of 8° C, a density of the sediment particles of 2.65 g · cm^{-3}, and on a flow time determined by the assumption that on an average $^1/_3$ of the lake volume is taking part in the through flow. These simplifying assumptions are probably rather valid when the dis-

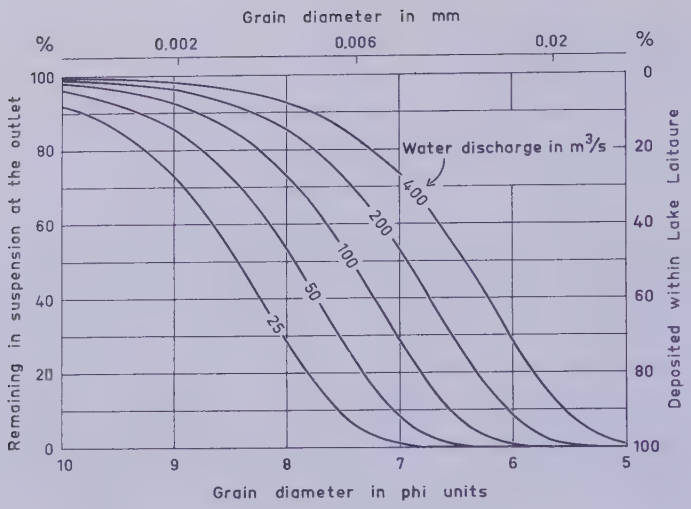


Fig. 69. Calculated amounts of sediment remaining in suspension at the outlet section of Lake Laitaure, or deposited within the lake upstream of this section, in percentage of the amounts discharged into the lake from the deltaic distributaries. The calculations are based on a water temperature of 8° C, a density of the sediment particles of 2.65 g · cm^{-3}, and that on an average $^1/_3$ of the lake volume is taking part in the through flow.

charge of suspended sediment is high enough to be of greater importance for the total annual sediment discharge.

According to the calculations the quantity of sand and coarse silt transported through the lake is rather insignificant in relation to that deposited, while the quantity of fine silt and clay transported through the lake often is greater than that deposited. The measurements of the suspended-sediment concentration, at the distributary mouths, are few, but indicate that the average concentration here is about 90 % of that at the proximal delta section when the sediment discharge is high. The concentration of fine silt and clay at the distributary mouths is probably about equal to that at the proximal delta section when the water discharge is lower than the bankfull discharge. Sediment discharges at the outlet section, estimated by means of Fig. 69, are of the same order as those estimated by the rating curve (Fig. 68), but gives somewhat lower discharges of coarser particles than expected on the basis of the grain-size analyses.

Chapter 10

MORPHOLOGY

The delta plain

The Laitaure delta is an area of subdued relief that sharply contrasts with the adjoining mountains which face their bare rock walls and steep talus slopes to the delta. Active and abandoned channels, levees and inter-levee basins constitute the major elements of the delta plain and surround islands and peninsulas of the former lake basin (see Plate I). As previously pointed out the boundary of the delta is determined by that of the original lake basin. Initial topographic irregularities influence splitting and closing of channels and formation of central and marginal delta basins, but in the main the deltaic pattern is the result of fluvial activity.

The subaerial part of the delta has a length of about 7 km. The width of the delta plain is about 2 km in the distal and middle parts but generally less than 1 km in the proximal part.

Channels

A branching channel pattern characterizes the delta plain but also large parts of the river course upstream of the delta. At the proximal boundary of the delta Rapaälven is divided into 3 larger branches, which join into a single channel upstream of the proximal delta section. The undivided reach is short. In the middle and distal parts of the delta plain the number of active and abandoned channels is comparatively great. The 3 major distributaries are called: Nuortätno (the north river), Kuoutätno (the middle river), and Årjeliätno (the south river).

The deltaic channels are shallow, especially in the distal part of the delta. Along most active channel reaches in the middle part of the delta average water depth is of the order of 2 m at bankfull stage but less than 1 m at mean water stage. In winter most of the smaller channels freeze to the bottom.

Pools are generally excavated where turbulence is heightened and are therefore often found at sharper channel bends, channel forks, and channel junctions. In 1958 the greatest channel depth was found at the surveyed channel junction (Fig. 80). At bankfull stage maximum depth here exceeded 4 m.

The levee-banks are cohesive and often very steep (Fig. 84). Along several channel reaches the width of the water surface is therefore rather constant as long as the water stage is higher than the bar crests but lower than bankfull. At low water the active channels have a somewhat braided appearance and several smaller channels are then closed at the head. The flow network at low water therefore differs considerably from that at high water.

The bed configuration varies rapidly at high-water discharges when the rate of bar migration is high. The variation of the average bed elevation during a flood passage, however, is probably relatively small. In 1957 the mean bed elevation at the proximal delta section was 15 cm lower on July 6 than on July 21 (see Fig. 70). This difference in mean bed elevation was due partly to scour and partly to bar migration during the passage of a flood. This flood was divided into 3 peaks. The highest peak occurred on July 17—18.

Boulders belonging to the till cover or released from adjoining rock walls influence the bed configuration at several localities. Other obstacles such as bushes, trees, and blocks of soil, that slide into the water on the erosion scarps, also affect the local bed configuration (Fig. 74). During falling stage, series of small terraces are often formed in the exposed bars by wave action. Small wind dunes are generally formed at low-water in autumn when large parts of the exposed channel beds dry up. The effect of ice on the bed configuration is most obvious just below the small rapids that mark the proximal boundary of the delta plain. At ice break-up, ice-jams are often formed below these rapids. In some years these ice-jams cause overtopping of adjoining levees.

On the whole the grain-size of the bed material decreases slowly in a downstream direction. Gravel bars are common just downstream of the proximal delta boundary. As exemplified by sample 1 in Fig. 75, these gravel bars are generally

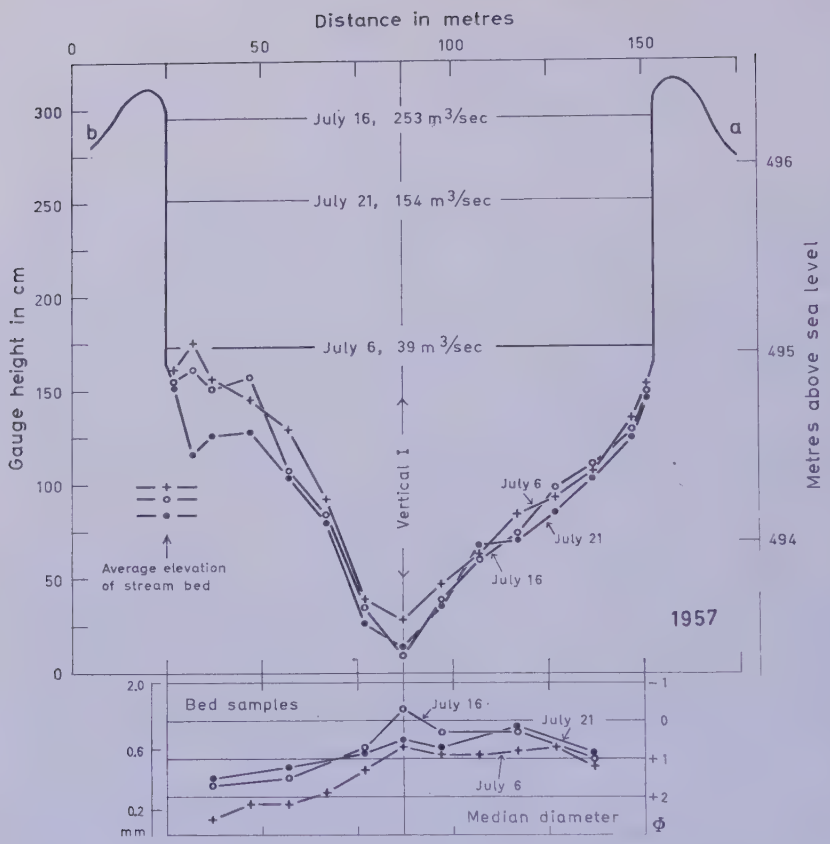


Fig. 70. Section a—b at the proximal part of the Laitaure delta. Channel cross sections and median particle sizes of bed materials during progress of flood, July 1957. Vertical exaggeration × 50.

characterized by a bimodal grain-size distribution with one maximum in the sand fraction and the other in the gravel fraction. Further downstream the migrating bars consist mainly of well-sorted sands, as exemplified in Fig. 75 by the grain-size distributions of samples 2 and 3. These samples are from the middle part of the delta; sample 2 from the proximal surface of a migrating transverse bar; sample 3 from the downstream crest of a point bar covered by current ripples.

The grain-size of the bed material varies across the channels, as illustrated in Fig. 70 by median grain sizes of collected bed samples, and in Figs. 38 and 77 by the grain sizes at which 10 %, 25 % (Q_1), 50 %, 75 % (Q_3) and 90 % of the bed samples (by weight) is finer. Normally the coarsest material is found in the deeper parts of the active channels. Erosional truncation and enrichment of coarser residual material takes place during rapidly rising water stages. The grain size of the bed surface therefore varies during a flood passage (see Fig. 70). In the abandoned channels the introduced fill mainly consists of silty sands.

Of special morphological interest are the areas where channels bifurcate and merge or re-join. The detail maps A and B (Figs. 76 and 80) are mainly based on surveyings by *M. Mattson*, who assisted the author during the field work in the summer of 1958.

The surveyed channel fork

Most of the channel forks are asymmetrical with one branch smaller than the other. The smaller branch often diverges at a greater angle from the trunk stream. Map A (Fig. 76) shows several morphological features that are common to many of the channel forks on the delta plain.

The resistance to flow caused by the bank at the point of bifurcation implies a damming action, flattening the slope of the water surface upstream and steepening it in a lateral and downstream direction. At high-water discharges

Fig. 71. View from Skerfe looking across the distal part of the Laitaure delta. The surveyed channel fork and the surveyed channel junction are shown in center of picture. July 30, 1959. Water discharge (Q_R): 50 m³/sec. Water stage in Lake Laitaure: 126 cm.

Fig. 72. The right branch channel at the surveyed channel fork, just downstream of section a—b (Fig. 76). July 6, 1958. Water discharge (Q_R): 65 m³/sec.

Fig. 73. A longitudinal bar in Nuortätno, northeast of Saivajaurats. Note the steeper side to the left, facing the adjoining levee bank, and the more gentle slope to the right. July 29, 1959. Water discharge (Q_R): 62 m³/sec.

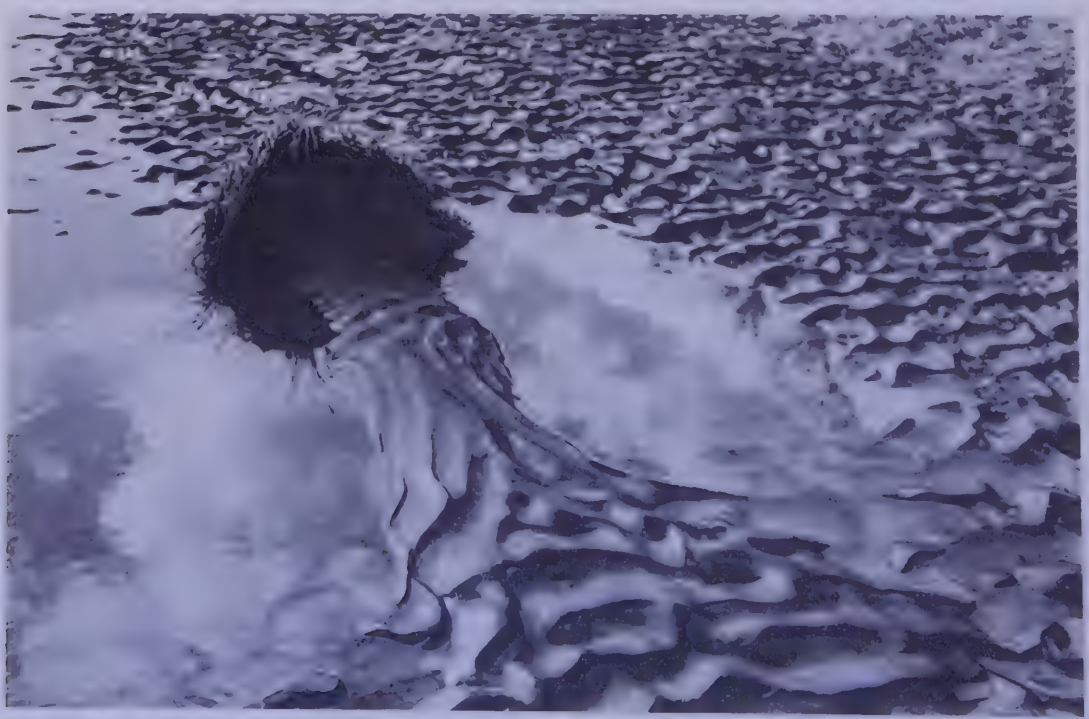


Fig. 74. Obstacles, such as blocks of soil, that slide into the water on the erosion scarps, affect the local bed configuration.

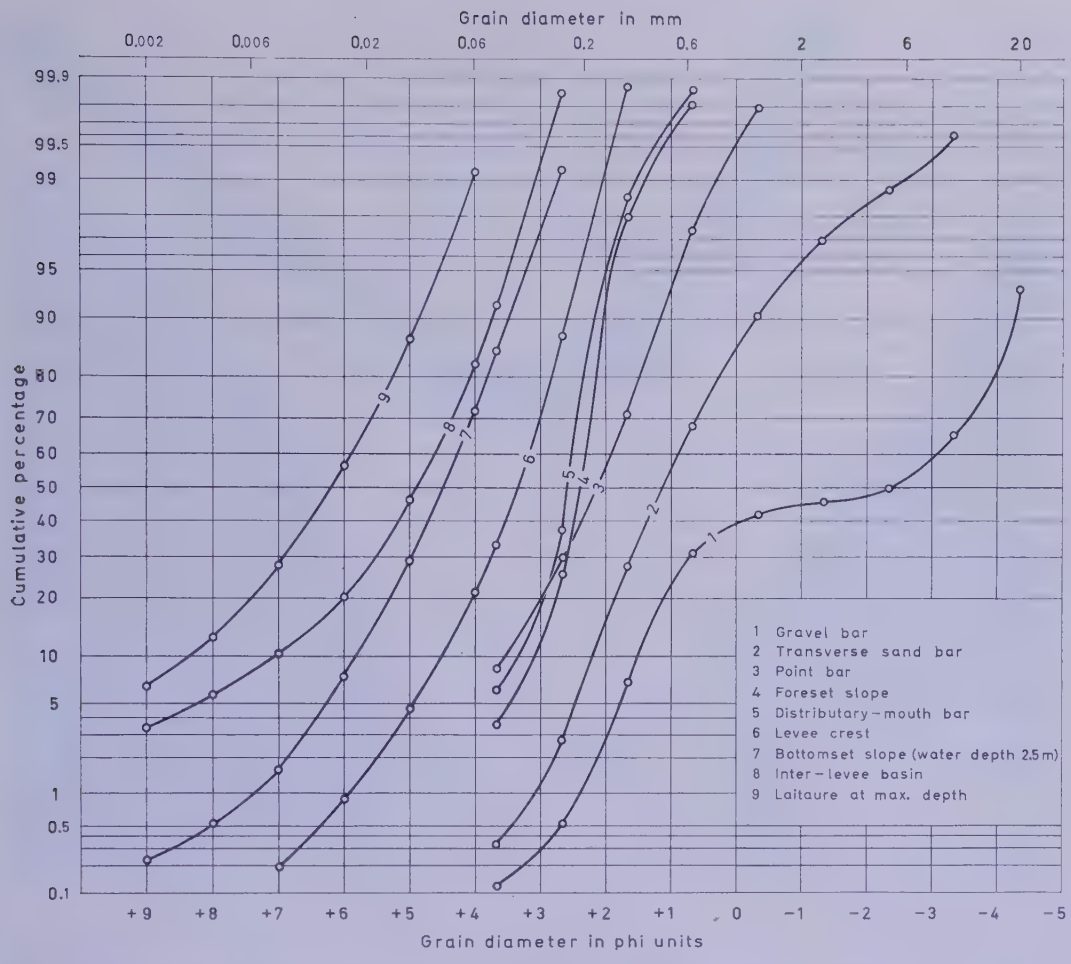


Fig. 75. Grain-size distributions of samples from the delta plain (1, 2, 3, 6, 8) and the delta front (4, 5, 7, 9).

the difference in water surface elevation between the point of bifurcation and the temporary gauge station at the entrance to the left channel may exceed 0.2 m. At high-water discharges strong eddies are formed at and just downstream of the point of bifurcation. Deep, narrow pools are here excavated in the channel beds. The adjoining levee-bank is undercut, and slides sometimes occur, but the rate of lateral erosion seems to be relatively slow.

A lobate bar, having a steep distal slope, is formed just upstream of the point of bifurcation. This bar partially blocks the entrance to the right channel. The shape of the bar varies somewhat with the flow conditions. At mean water stage, when part of the bar is emerged, the flow into the right channel is split into two currents. At low water the flow into the right channel follows the narrow passage between the distal slope of the bar and the undercut levee-bank.

The flow velocities are normally asymmetrically distributed at the entrance to both the right and left channel. At high-water discharges reverse eddies are formed at the upstream part of the entrance to the branch channels. These eddies reduce the effective areas of the cross-sections and accelerate the constricted flows. As exemplified by Fig. 77, the asymmetry of the channel cross-sections is very pronounced at the entrance to the right channel (see also Fig. 72). The three curving levee ridges, formed along the inside of the entrance to this channel, mark the lateral displacement of the channel entrance.

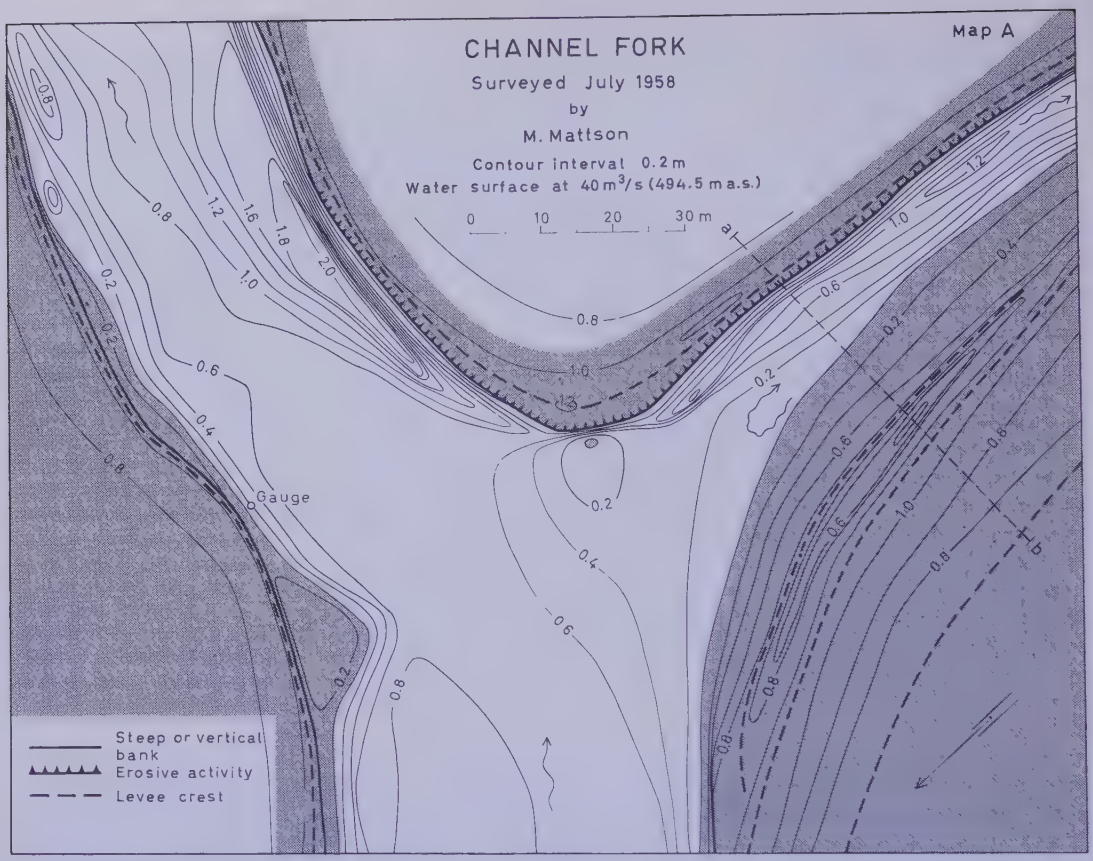


Fig. 76. Map of channel fork. For location of map A, see Plate I.

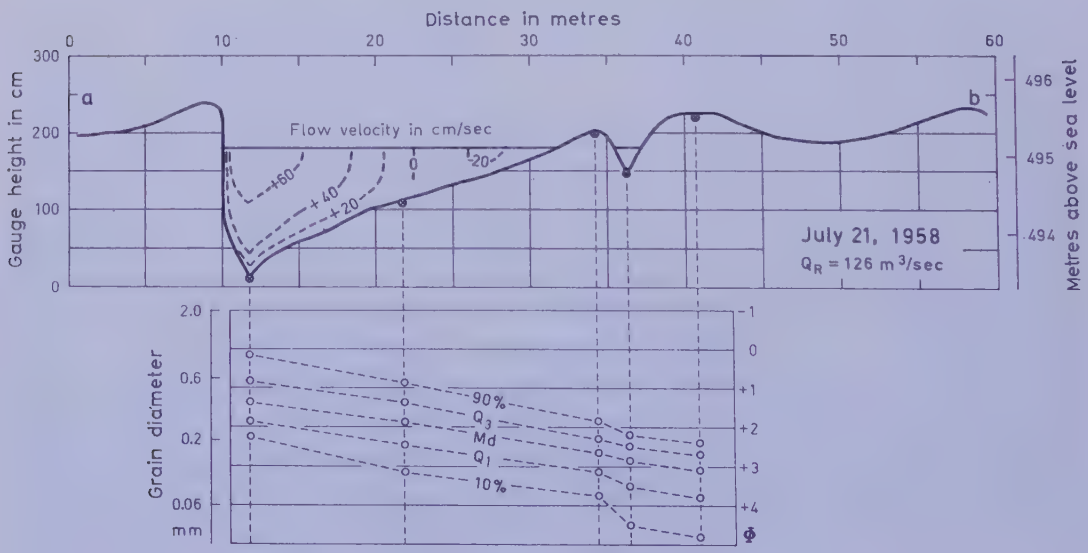


Fig. 77. Section a—b at the channel fork (Fig. 76). Vertical exaggeration × 5.

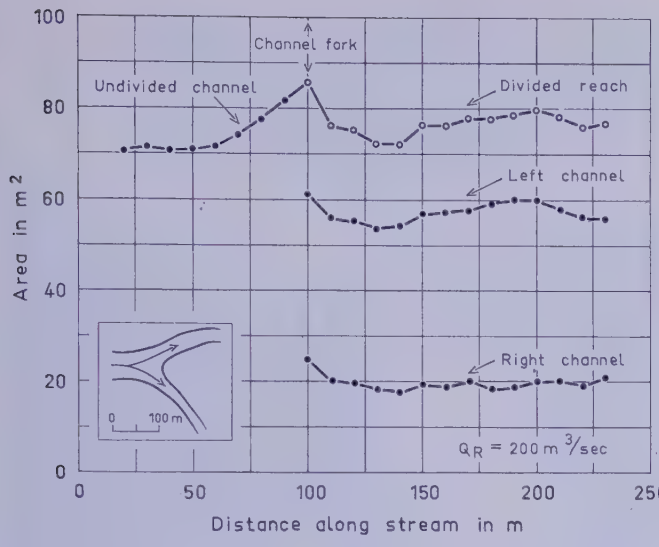


Fig. 78. Variation of cross-sectional area along the channels at the channel fork (Fig. 76).

The values of cross-sectional area and mean depth, plotted in Figs. 78 and 79, are referred to the water stage that normally prevails at the channel fork when the water discharge is 200 m^3/ sec. at the proximal delta section. As shown by Fig. 78 the cross-sectional area of the divided reach is larger than that of the undivided channel, but the difference is rather insignificant. The right channel is considerably smaller than the left channel. At high water the cross-sectional area of the right channel is about 1/3 that of the left channel.

The variation of mean depth along the surveyed parts of the channels is illustrated in Fig. 79. The mean depth of the divided reach was computed by dividing the cross-sectional area of flowing water by the total width of water surface in the two channels. The channels are shallowest at the channel fork. The left channel is not only deeper than the right channel but also deeper

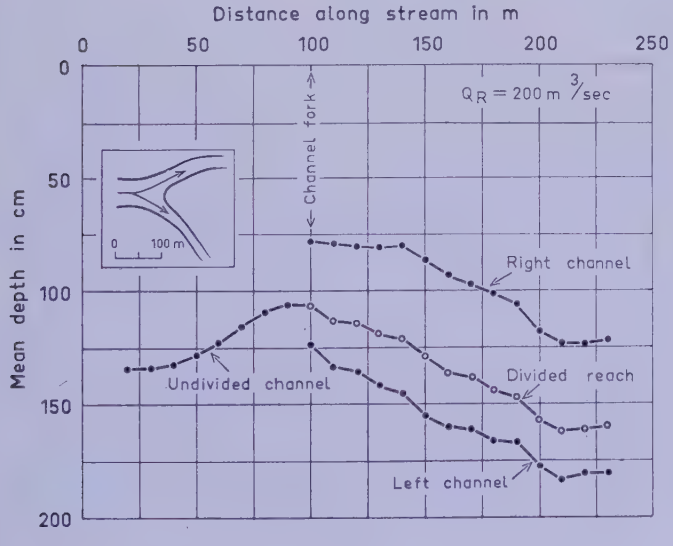


Fig. 79. Variation of mean depth along the channels at the channel fork (Fig. 76).

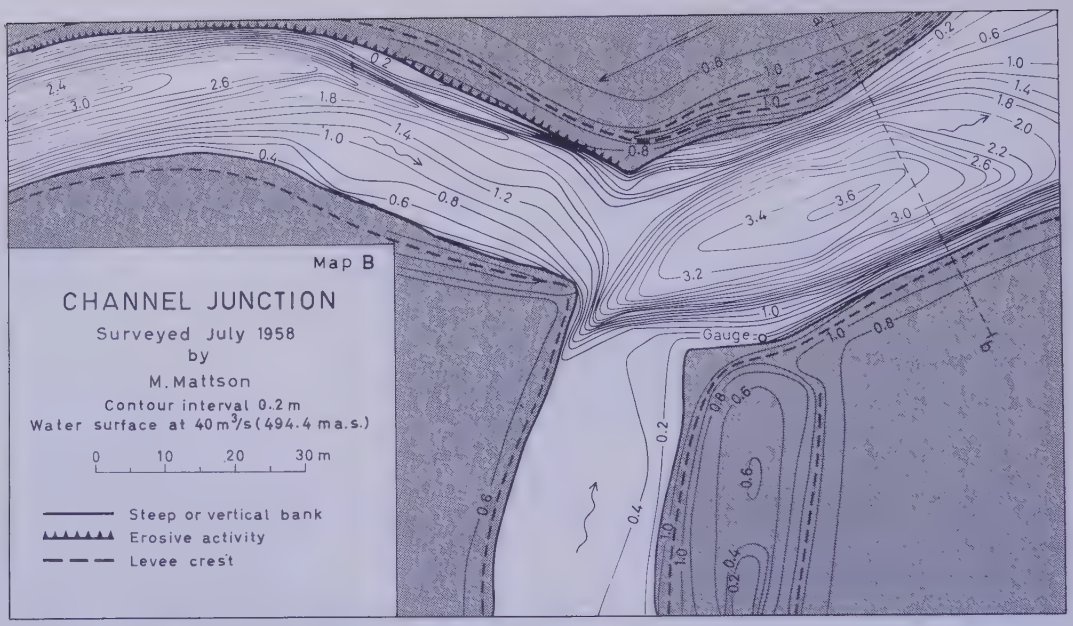


Fig. 80. Map of channel junction. For location of map B, see Plate I.

than the trunk stream. This may indicate that the efficiency of this channel has increased in relation to that of the right channel, and that vertical erosion has taken place in the channel as a result of altered equilibrium conditions.

The surveyed channel junction

A deep pool is generally excavated in the bed at the place where two channels merge or rejoin at a sharp angle. At the surveyed channel junction (Fig. 80) a very large pool is formed where the two branch channels meet. The pool occupies practically the whole width of the main channel.

Where the currents meet the flow is accelerated, turbulence is heightened, and the turbulent structure of the flow is here very complicated. An example of the transverse variation of the longitudinal velocity component is given in Fig. 81.

Due to erosional truncation within the pool area the upper parts of the older delta front deposits have been eroded and partly replaced by coarser channel lag deposits. The sandy channel bed is comparatively thin in the section where core samples were taken (see Fig. 81). It is therefore probable that the underlying lake deposits still may be eroded when the water discharge is rapidly rising.

As shown by Figs. 82 and 83 the cross-sectional area and the mean depth vary considerably along the channels at the channel junction. The left channel is considerably larger and deeper than the right channel. The variation of mean depth in the left channel is partly due to the pool that is excavated in the bed at the bend of the channel.

Levees and basins

The channels are generally confined by well developed levees, often having comparatively steep backslopes. At present and former accumulation sides the levees are often divided into several, more or less parallel, usually closely spaced ridges that mark the lateral displacement of the adjoining channel. The levee-banks are generally very steep at the outside of the channel bends and at the downstream part of the channel forks, where slides sometimes occur at the erosion scarps.

On the whole the levees decrease in height towards the channel mouths. The average longitudinal slope of the delta plain, as defined by the average slope of the levee crests along the active channels, is about 0.2 per mille in the middle part of the delta, and about 0.3 per mille in the proximal part. In the distal part of the

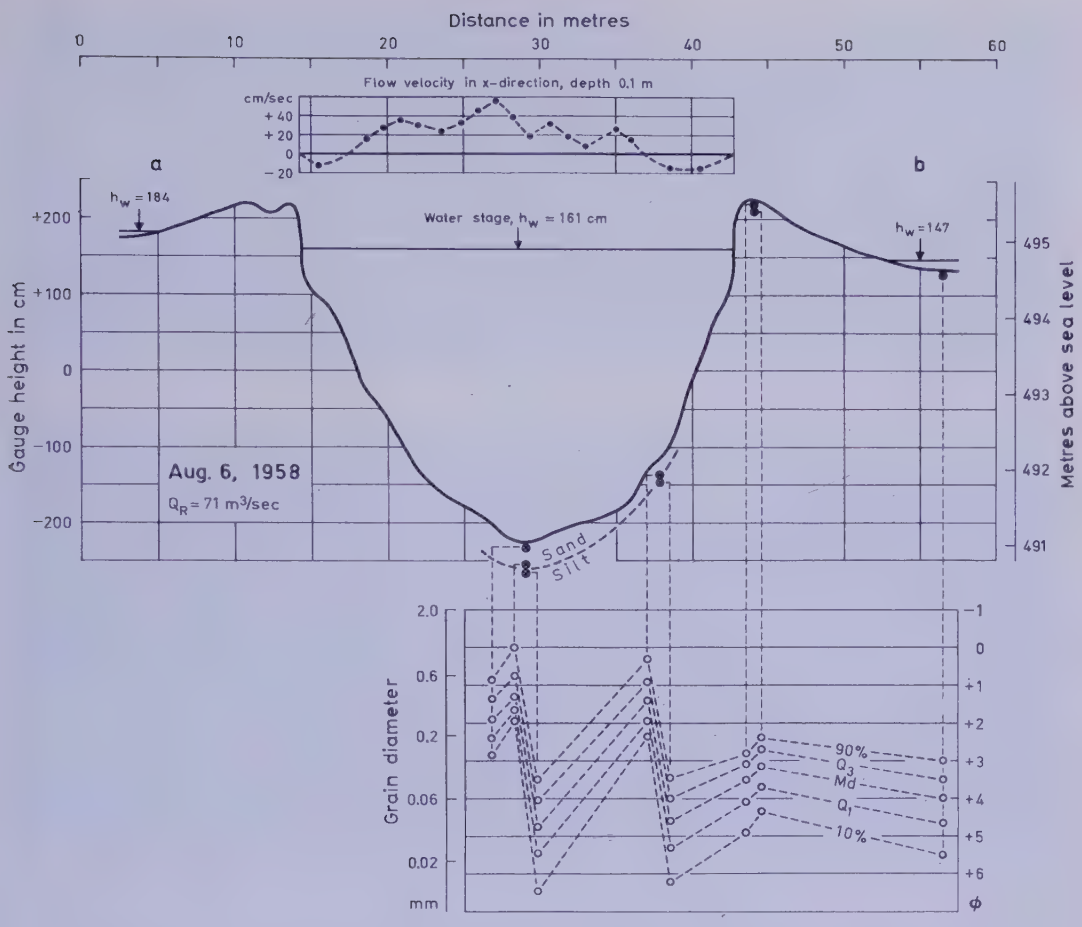


Fig. 81. Section a—b at the channel junction (Fig. 80). Vertical exaggeration × 5.

delta the levees are flooded more frequently and during longer periods than further upstream. At the proximal part of the delta they are topped by flood-water only about 2 out of 3 years on the average.

The birch forest, that is found on the highest levee crests in the proximal and middle parts of the delta, is generally rich in luxurious tall herbs. Large parts of the levees are generally occupied by dense willow thickets, mainly consisting of *Salix glauca*, *S. lanata*, *S. lapponum*, and *S. phylicifolia*, with *S. lapponum* and *S. phylicifolia* on the lower levels. Towards the delta lakes willow thickets generally give way to more or less extensive sedge fens, often dominated by *Carex aquatilis*, and sometimes followed by *Equisetum fluviatile*. On the rapidly growing, distal parts of the delta the pioneer vegetation consists of *Deschampsia alpina*, often followed by *Equisetum arvense f. alpestre*, *Calamagrostis neglecta*, *Carex aquatilis*, *Juncus arcticus*, and *Eriophorum Scheuchzeri* (see further *Tengwall*, 1925).

The larger inter-levee basins are subdivided by minor levee complexes into smaller lake basins. Marginal delta lakes are common where the original shore configuration is irregular. The delta lakes are shallow, and many of them freeze to the bottom in winter. Several of the lakes are partly filled by miniature deltas, formed at the mouths of inflowing channels.

In those delta lakes that have only one channel connection with the active distributaries the connecting channel generally carries turbid river water into the lake during rising river stages and

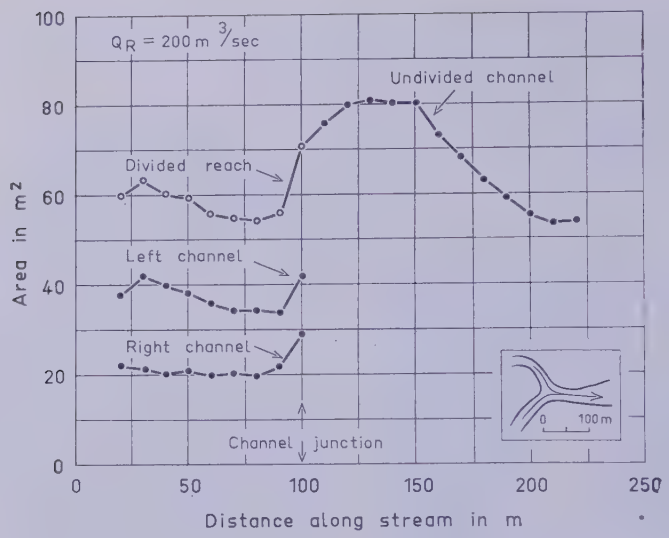


Fig. 82. Variation of cross-sectional area along the channels at the channel junction (Fig. 80).

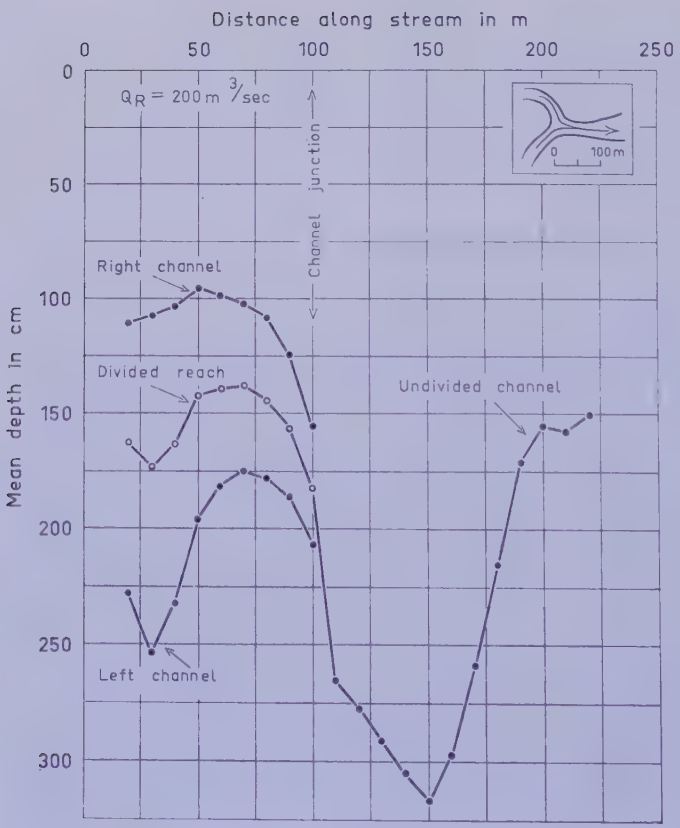


Fig. 83. Variation of mean depth along the channels at the channel junction (Fig. 80).

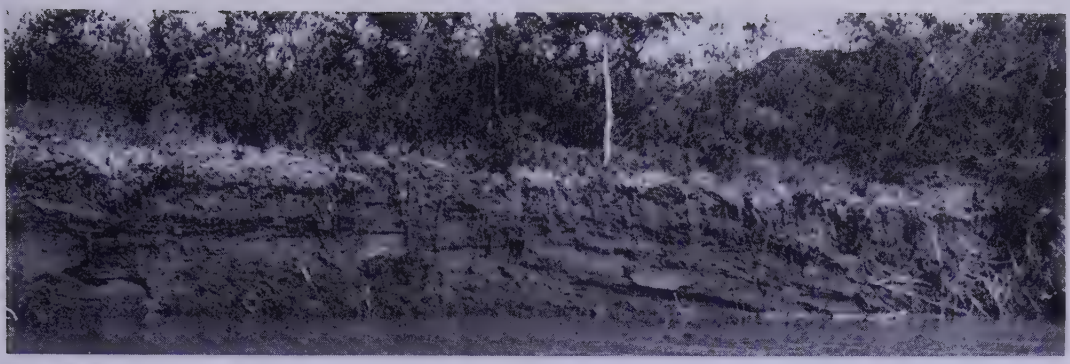


Fig. 84. Levee-bank close to the proximal boundary of the Laitaure delta. August 9, 1959. Water discharge (Q_R): 36 m^3/sec.

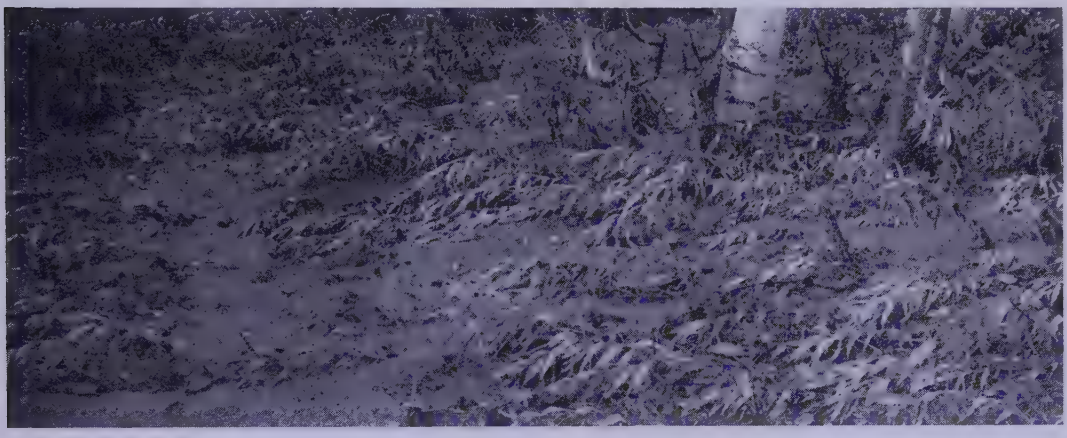


Fig. 85. Part of levee crest just upstream of the proximal boundary of the Laitaure delta. The vegetation is partly buried by a sandy sediment layer, 2—3 cm thick, that was deposited during a flood in August 1954.

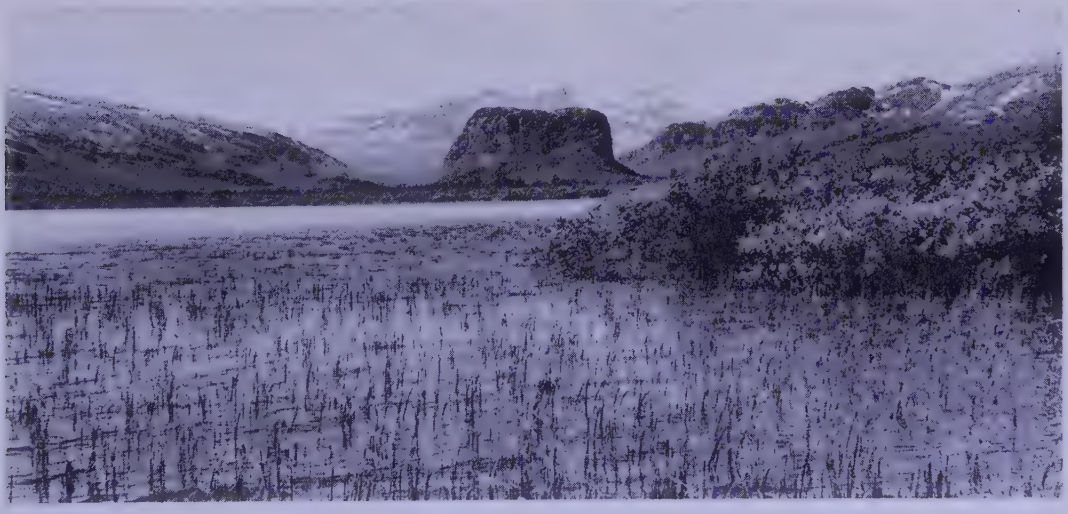


Fig. 86. Part of the distal inter-levee basin west of the surveyed channel junction. July 4, 1958. Water stage in Lake Laitaure: 184 cm.

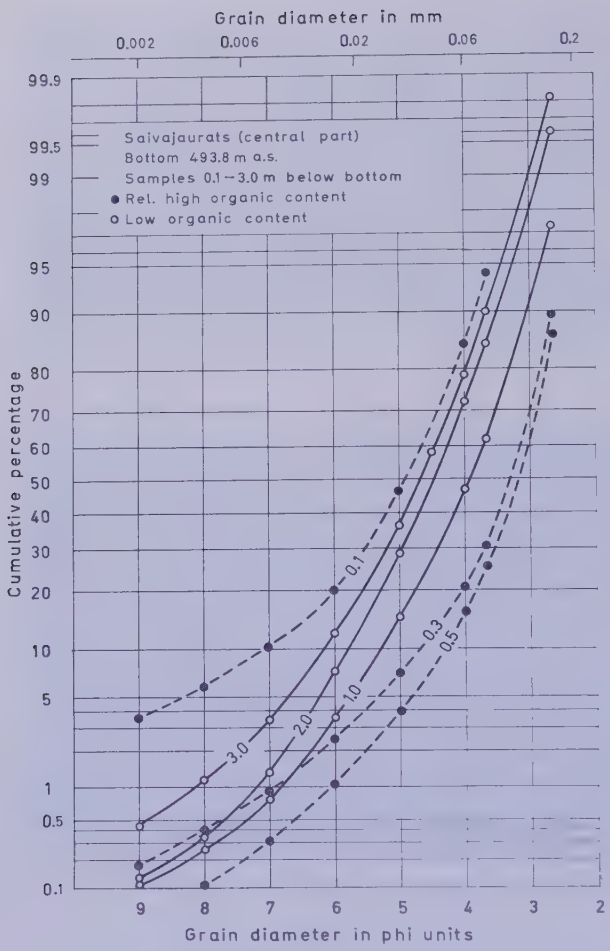


Fig. 87. Grain-size distributions of samples taken 0.1—3.0 m below the bottom of the delta lake Saivajaurats. For location of the lake, see Plate I.

clearer lake water out to the distributaries during falling river stages. Several of the delta lakes are entirely cut off from channel connection with the distributaries. In these lakes the water stage may differ considerably from that in the distributaries. In some of these lakes the water stage may sometimes be more than 1 m higher or lower than in the adjoining distributaries.

The grain size of suspended sediment deposited from floodwaters decreases with increasing distance from the active distributaries. This is in Fig. 75 illustrated by the grain-size distributions of the samples 6 and 8 of suspended sediment deposited during a flood in July 1957. Sample 8 is from the central part of the marginal delta lake Tjappesjaurats (for location, see Plate I), where a layer, about 1 mm thick, was deposited. The rate of deposition here was measured by means of a specially designed apparatus (for description of the apparatus, see *Axelsson*, 1955). Sample 6 is from the adjoining levee crest where a layer, about 2 cm thick, was deposited.

Generally the levees are characterized by alternating, often somewhat distorted, coarser and finer layers consisting mostly of fine sand and coarse silt (see Figs. 77 and 81). Within those inter-levee basins that lack channel connection with the active distributaries the surface deposits are generally indistinctly layered and often have a comparatively high content of organic material.

The vertical distribution of sediments is exemplified in Fig. 87 by grain-size distributions of core samples from the central part of the

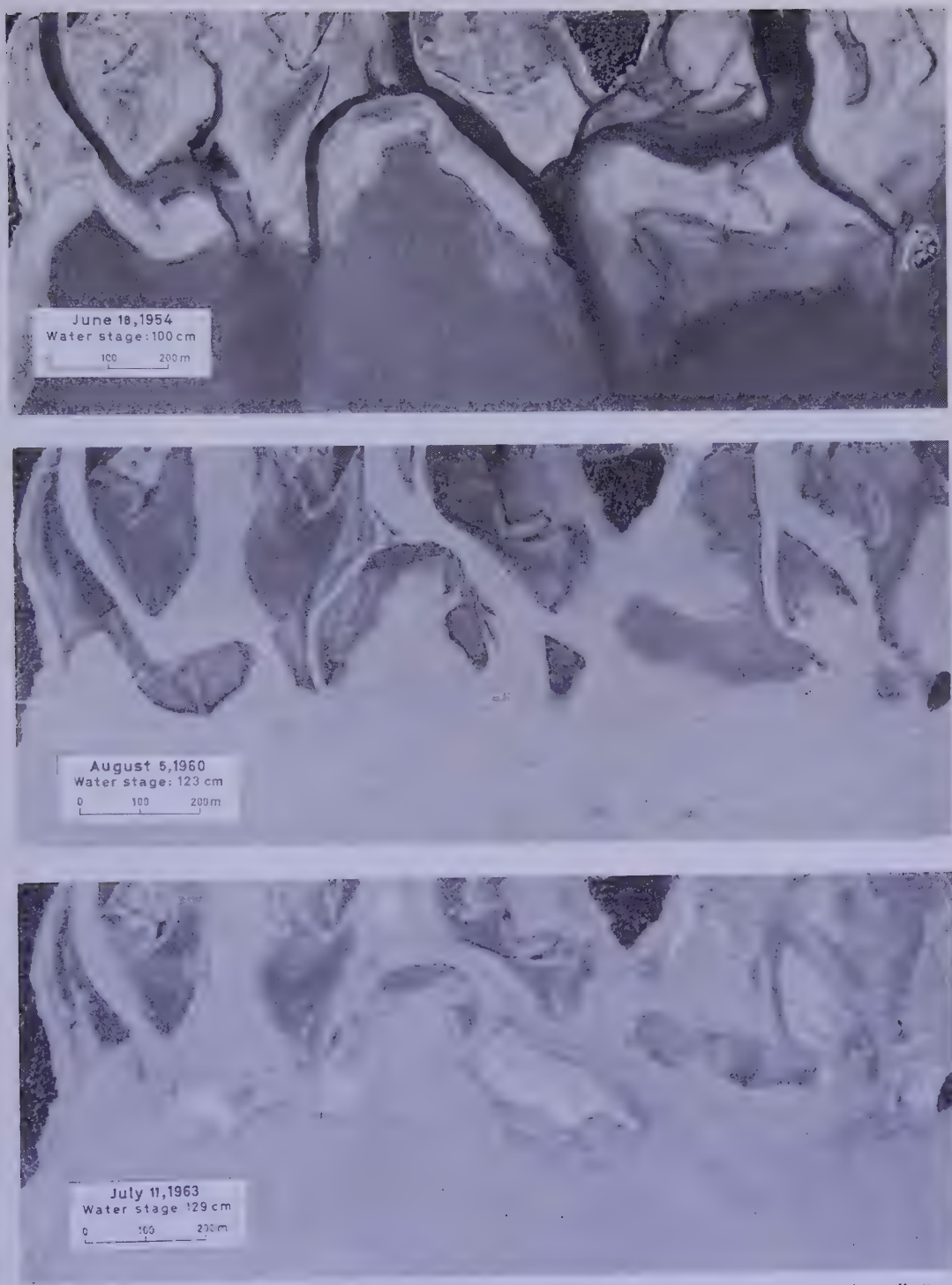


Fig. 88. The middle and southern parts of the delta front in 1954, 1960 and 1963. Air photographs by *Rikets allmänna kartverk*.

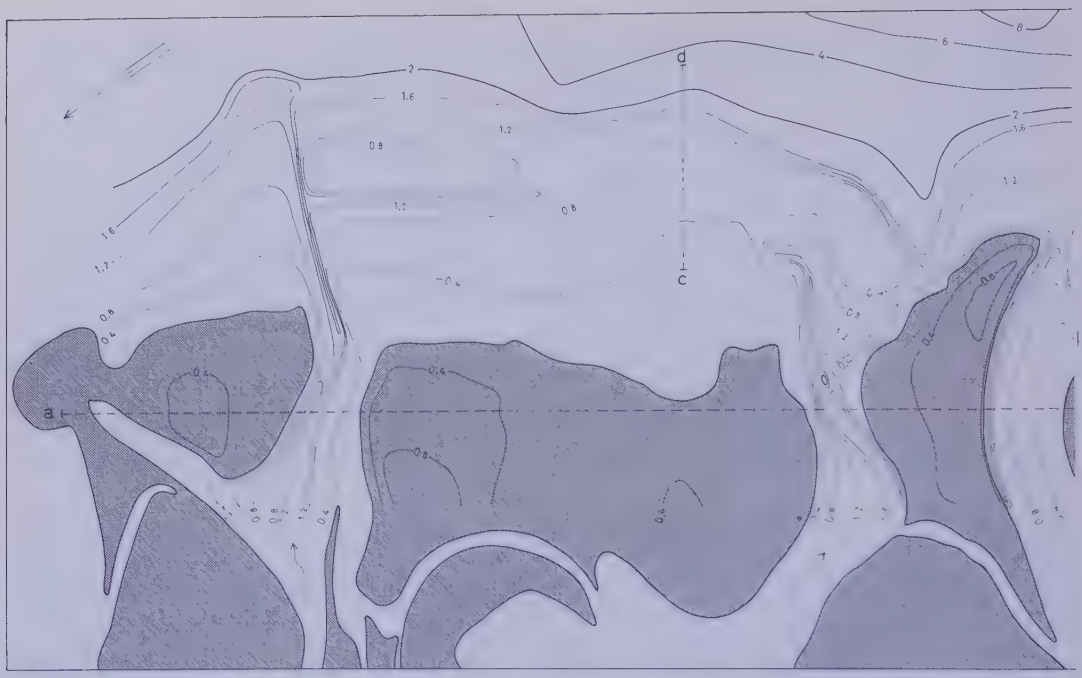


Fig. 89. Map of the middle and southern parts of the delta front in Lake Laitaure. For location of map C, see Plate II.

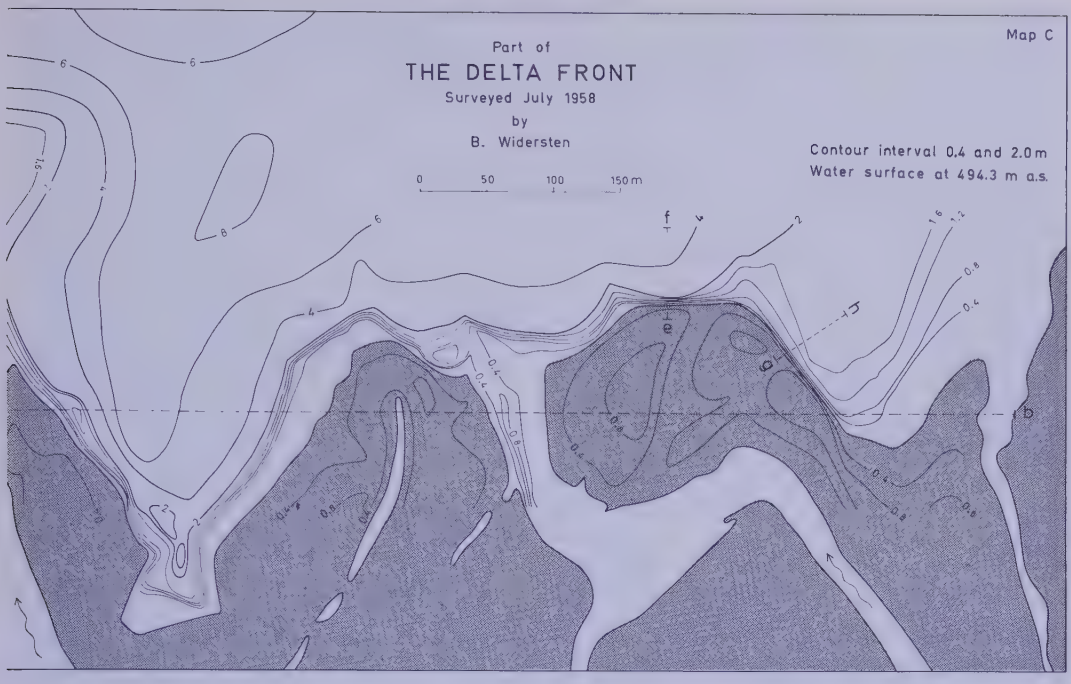
delta lake Saivajaurats, taken 0.1—3.0 m below the bottom of the lake. The coarsest delta sediments are here found about 0.5 m below the lake bottom. These sediments were probably deposited in a shallow inter-distributary bay. The uppermost sediments are deposited in a more sheltered environment and have a comparatively high content of organic material. The grain-size distributions of the deeper core samples correspond to those of sediments deposited in Laitaure in front of the present distributary mouths.

The delta front

At the distributary mouths the flow pattern generally simulates that of expanding jets with complementary zones of reverse flow, often characterized by appreciable circulation and large-scale eddying. The lateral zones of flow separation and backflow vary with the variation in water level. Flow expansion is often restricted by density differences. In summer the inflowing river water is normally denser than the surface water in Laitaure, and density currents of the underflow type are often formed at the delta front during the periods when the sediment discharge is high.

The boundary, where turbid river water dives below clearer lake water, may be very sharp during rapidly rising river stages, and is often marked by floating debris. The density contrast between the inflowing river water and the lake water was measured on several occasions at the plunge point. It was found that the density contrast here could be as high as 0.0008 and that underflows still were formed when the density contrast was as low as 0.0003. At high-water discharges the plunge-point was situated a considerably greater distance from the distributary mouths at the northern part of the delta front, where the lake is very shallow, than at the southern part, where the lake is deeper, and the delta-front slopes steeper.

At the delta front the levees are low and poorly developed, the channels are poorly defined and bar deposits of active and abandoned channels merge laterally to form greater distal lobes. Along the shallow northern part of the delta front channel deepening is greatly hindered by erosion-resistant material. The southern part of the delta front is on the whole built out into deeper



water, but also here is channel formation and extension partly influenced by the intial topography of the frontal lake basin.

The detail map of the middle and southern parts of the delta front (Map C, Fig. 89) is mainly based on surveyings by *B. Widersten*, who assisted the author during the field work in the summer of 1958. As shown by the contour lines the tendency to blocking, splitting, and lateral migration of the channels is obvious. The bar and channel pattern therefore varies rapidly with time, as illustrated by the restituted air photographs (Fig. 88).

The frontal bars are built up of easily eroded material, and their shape is therefore sometimes modified by wave action (see Fig. 94). In general, however, the energy of the waves is low at the delta front, and on the whole the direct morphological action of the shore processes is comparatively weak.

In 1958 the highest parts of the frontal bars or levees (partly stabilized by vegetation) reached about 1.8 m above the selected zero level (see Figs. 89 and 90) and about 0.9 m above the mean water level (0.9 m). The total duration of water stages in Laitaure higher than 1.8 m was in 1957 about 11 days and in 1958 about 13 days (Fig. 35). The calculated mean annual extreme water stages amount to 2.1 m and 0.5 m respectively.

Sediments with median diameter of about 0.18 mm are the best sorted and are often characteristic of river-mouth bars and deltaic foreset slopes. This is the case at the delta front in Laitaure as illustrated by the grain-size distributions of samples from the sediment surface (Figs. 75, 90 and 91). The sandy bar deposits are generally capped by somewhat finer-grained material in those areas where they are colonized by vegetation. The inter-distributary bay deposits are less well sorted than the bar deposits. Silty deposits characterize the more sheltered parts of the inter-distributary bays and the deeper parts of the frontal slopes.

The inclination of the frontal slopes is greatly influenced by the original topography of the frontal lake basin. In the northern and middle parts of the delta front the inclination of the delta-front slopes is very low. Steeply inclined delta-front slopes of a real foreset character are only found at the southern part of the delta front. The dip angle of the steeper parts of the higher foreset slopes is here about 25°. The

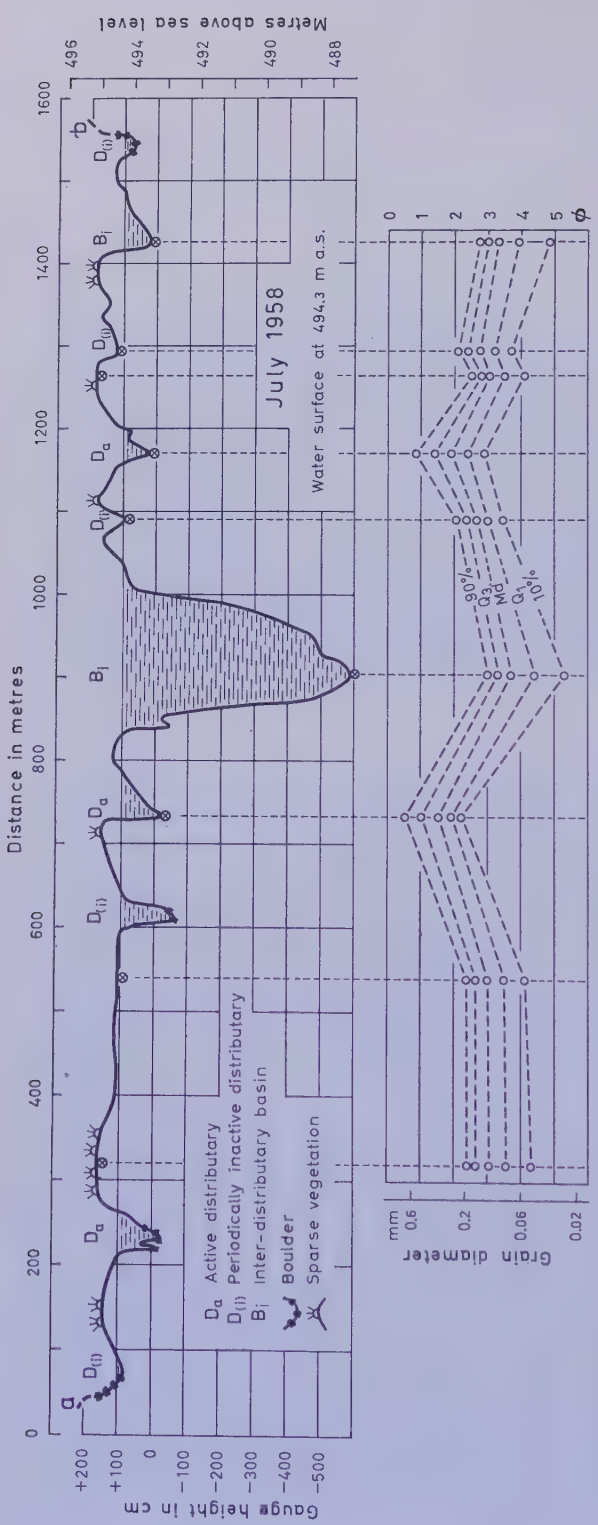


Fig. 90. Section a—b at the delta front (Fig. 89). Vertical exaggeration × 40.

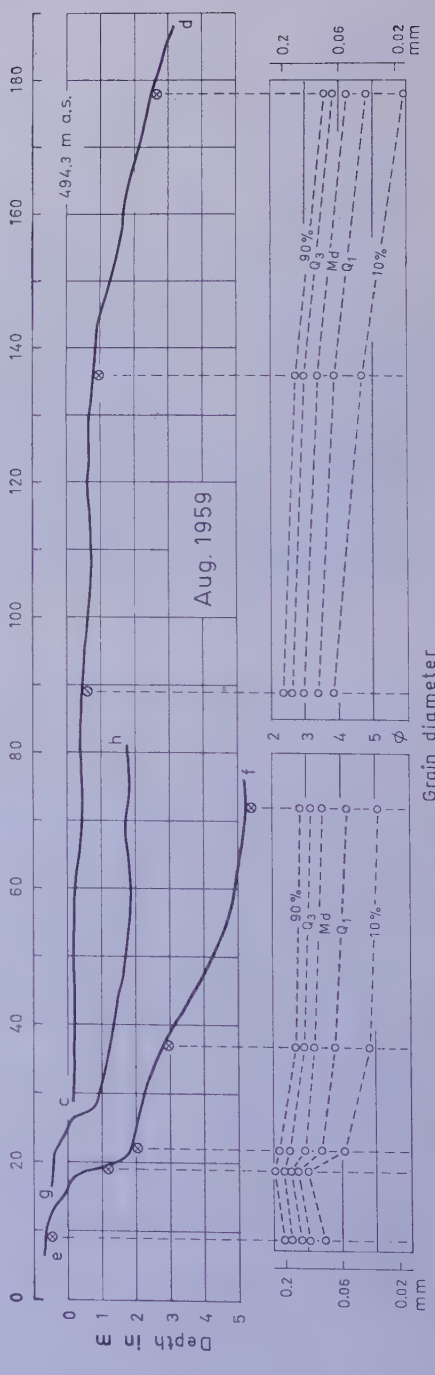


Fig. 91. The sections c—d, e—f, and g—h at the delta front (Fig. 89). Vertical exaggeration × 5.

Fig. 92. View from Tjakkeli looking along the delta front in Lake Laitaure. August 19, 1954. Water stage: 139 cm.

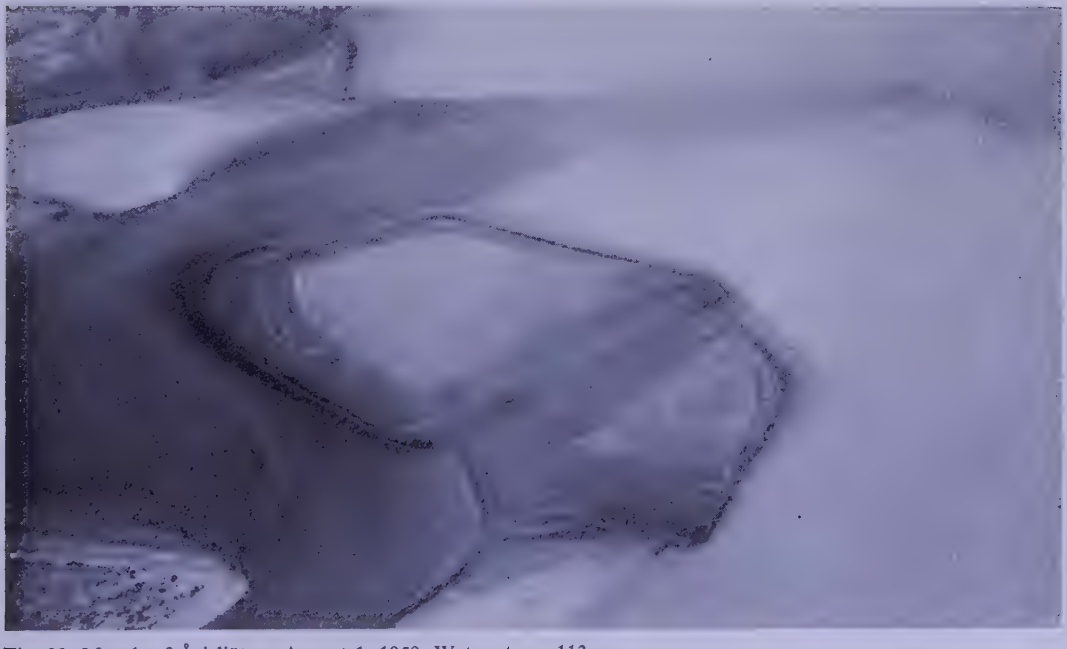


Fig. 93. Mouth of Årjeliätno. August 1, 1959. Water stage: 113 cm.

Fig. 94. Part of the distal lobe at the mouth of Årjeliätno. The maximum height of the wave-cut scarp is 11 cm. August 1, 1959. Water stage 112 cm.

Fig. 95. Inter-distributary bay at the southern part of the delta front, southwest of the distal lobe (Fig. 94). August 1, 1959. Water stage: 112 cm.

Fig. 96. Driftwood logs on the bars often lodge with the roots pointing upstream, the trunk downstream. August 1, 1959. Water stage: 112 cm.

Fig. 97. Pioneer vegetation (*Deschampsia alpina*) in the middle part of the delta front. July 18, 1958. Water stage: 98 cm.

Fig. 98. Marginal, periodically inactive, distributary mouth at the southern boundary of the delta front. August 1, 1959. Water stage: 112 cm.

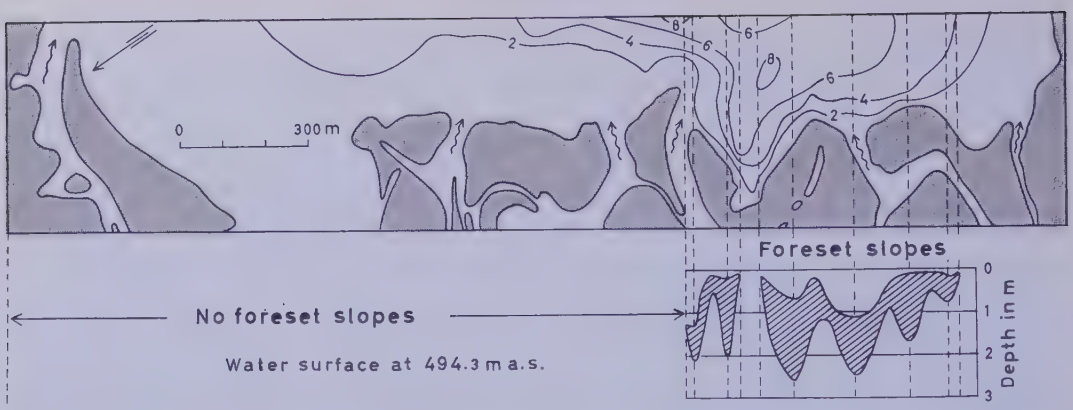


Fig. 99. Vertical and lateral extent of foreset slopes (frontal slopes steeper than 10°) at the delta front in Lake Laitaure (marked by striated areas). July 1958. Vertical exaggeration × 100.

maximum height of the foreset slopes was in 1958 about 2 m. The observed basal contacts were tangential.

The vertical and lateral extent of foreset slopes (frontal slopes of aggradational origin steeper than 10°) is shown in Fig. 99. The depth to the upper and lower boundaries of the foreset slopes (as defined above) varies along the delta front and reflects the position of active and abandoned distributary mouths. The depth of distributary mouths bears a direct relation to distributary discharge and they tend to choke up when part of their flow is diverted. The depth to the upper boundary of the foreset slopes is therefore greater at the mouths of the larger, active distributaries than at the mouths of the smaller, periodically inactive or abandoned distributaries.

The depth to the lower boundary of the foreset slopes is more influenced by the original topography of the frontal lake basin but reflects also the position and the activity of the distributary mouths. At those parts of the delta front that in 1958 had a marginal position in relation to the distributary mouths or the more active bedload paths the depth to the lower foreset boundary was generally small.

The vertical and lateral extent of foreset slopes is small at the present delta front in Laitaure. This means that the extent of steeply inclined foreset beds also is small in the younger parts of the delta. The grain-size distributions of core samples from the older parts of the delta indicate that also here the extent of foreset beds is small.

As the newly formed channels migrate laterally, bifurcate, and extend basinward, upper parts of underlying delta front deposits will be eroded and partly replaced by channel fill deposits, levees will be formed along the channel margins, and inter-levee basins will be embayed and later cut off from the frontal lake basin. The rate of advance of the Laitaure delta is high as compared to that of most lake deltas in Sweden. If we assume that the amount of sediment transported to the Laitaure delta will be of the same order in the future as at present the remaining part of Laitaure will be filled up by sediments within a period of 1,000 years.

LIST OF SYMBOLS

A	Area
a	Distance above the bed
b	Width
C	An empirical constant (different value for different equations); Chezy coefficient
c	Suspended sediment concentration (in dry weight per unit of volume)
c_a	Concentration of sediment of a given grain size at a distance a above the bed
c_s	Sediment concentration near the water surface (5 cm below the water surface)
c_y	Concentration of sediment of a given grain size at a distance y above the bed
$\tilde{c}$	Weighted mean sediment concentration
$\tilde{c}_p$	Weighted mean sediment concentration in percentage of the sediment concentration near the water surface
$\tilde{c}_{pf}$	Weighted mean sediment concentration in percentage of the sediment concentration near the water surface during falling stages
$\tilde{c}_{pr}$	Weighted mean sediment concentration in percentage of the sediment concentration near the water surface during rising stages
D	Depth; diameter
D_m	Mean depth
d	Grain size
d_e	Effective density
d_{65}	Grain size at which 65 per cent of the bed material (by weight) is finer
e	Base of Napierian logarithms
F	A function
g	Acceleration due to gravity
h	Height
h_o	Oil stage in the gauge pipe at Litnok
h_w	Water stage
K	Frequency factor
k	Von Kármán "constant"
k_s	A length parameter
L_b	Bed-load discharge in the river Rapaälven at the proximal delta section
L_{bf}	Bed-load feed
L_{bl}	Bed-load discharge in left channel
L_{br}	Bed-load discharge in right channel
L_{bt}	Total bed-load discharge
L_s	Suspended-sediment discharge
l	A distance
l_b	Bed-load discharge in weight per unit of time and width ($g \cdot cm^{-1} \cdot sec^{-1}$)
MHW	Mean annual extreme high-water stage
MLW	Mean annual extreme low-water stage
MW	Mean annual water stage
Md	Median grain diameter
m	Rank
m'	Immersed weight of a sediment particle
n	Number
P	Wetted perimeter
Q_B	Water discharge in the river Blackälven at the gauge station Tjåmotis 1
Q_L	Water discharge at the outlet of Lake Laitaure
Q_R	Water discharge in the river Rapaälven at Litnok or at the proximal part of the Laitaure delta
Q_l	Water discharge in left channel
Q_r	Water discharge in right channel
Q_t	Total water discharge
Q_1	First quartile; grain size at which 25 per cent of the material (by weight) is finer
Q_3	Third quartile; grain size at which 75 per cent of the material (by weight) is finer
R	Hydraulic radius
R_d	Hydraulic radius of a density current
Re	Reynolds number
r	Radius (of curvature)
r_c	Centerline radius
r_i	Inner radius
r_o	Outer radius
S	Slope angle
S_e	Energy gradient
S_w	Slope of the water surface
T	Temperature in degrees Celsius
T_r	Return period or recurrence interval
t	Time
u	Flow velocity
u_{max}	Maximum flow velocity in cross section
u_p	Flow velocity in percentage of the mean flow velocity in vertical
u_z	Transverse (radial) flow velocity
u_*	Shear velocity
$\bar{u}$	Mean flow velocity in vertical
$\bar{u}_s$	Mean flow velocity in cross section
v_s	Settling velocity of a sediment particle
x	A longitudinal distance; a variable
$\bar{x}$	Mean of x
y	A vertical distance; distance above the bed; a reduced variate in frequency analysis
y_n	A statistical factor in frequency analysis
y_r	Relative distance above the bed
z	A lateral distance; the average slope of the log-log plot of $(D—y)/y$ versus the total concentration of suspended sediment
z_t	Theoretical exponent of the vertical distribution of suspended sediment of a given grain size
z_1	Actual exponent of the vertical distribution of suspended sediment of a given grain size or grain-size fraction
α	A correction factor
β	Ratio of the transfer coefficient for sediment to the momentum transfer coefficient
γ	A factor which indicates the proportion of the bed taking the fluid shear

δ	Thickness of the laminar sublayer
ϱ_f	Density of the fluid
ϱ_s	Density of the sediment particles
σ	Standard deviation
τ_c	Critical boundary shear
τ_o	Boundary shear
Φ	Probability or error integral; phi unit; a function
φ	Deviation angle

REFERENCES

Abbreviations of the names of journals and serial publications are mainly according to the system recommended by *Unesco*.

Special abbreviations

GFF Geologiska Föreningens i Stockholm förhandlingar (Transactions of the Geological Society of Sweden), Stockholm.

GIU Geografiska institutionen vid Uppsala universitet (Department of Geography, University of Uppsala), Uppsala.

SGU Sveriges Geologiska Undersökning (Geological Survey of Sweden), Stockholm.

SMHA Statens Meteorologisk-hydrografiska anstalt (The Swedish Meteorological and Hydrological Institute, before 1946), Stockholm.

SMHI Sveriges Meteorologiska och Hydrologiska Institut (The Swedish Meteorological and Hydrological Institute, from 1946), Stockholm.

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